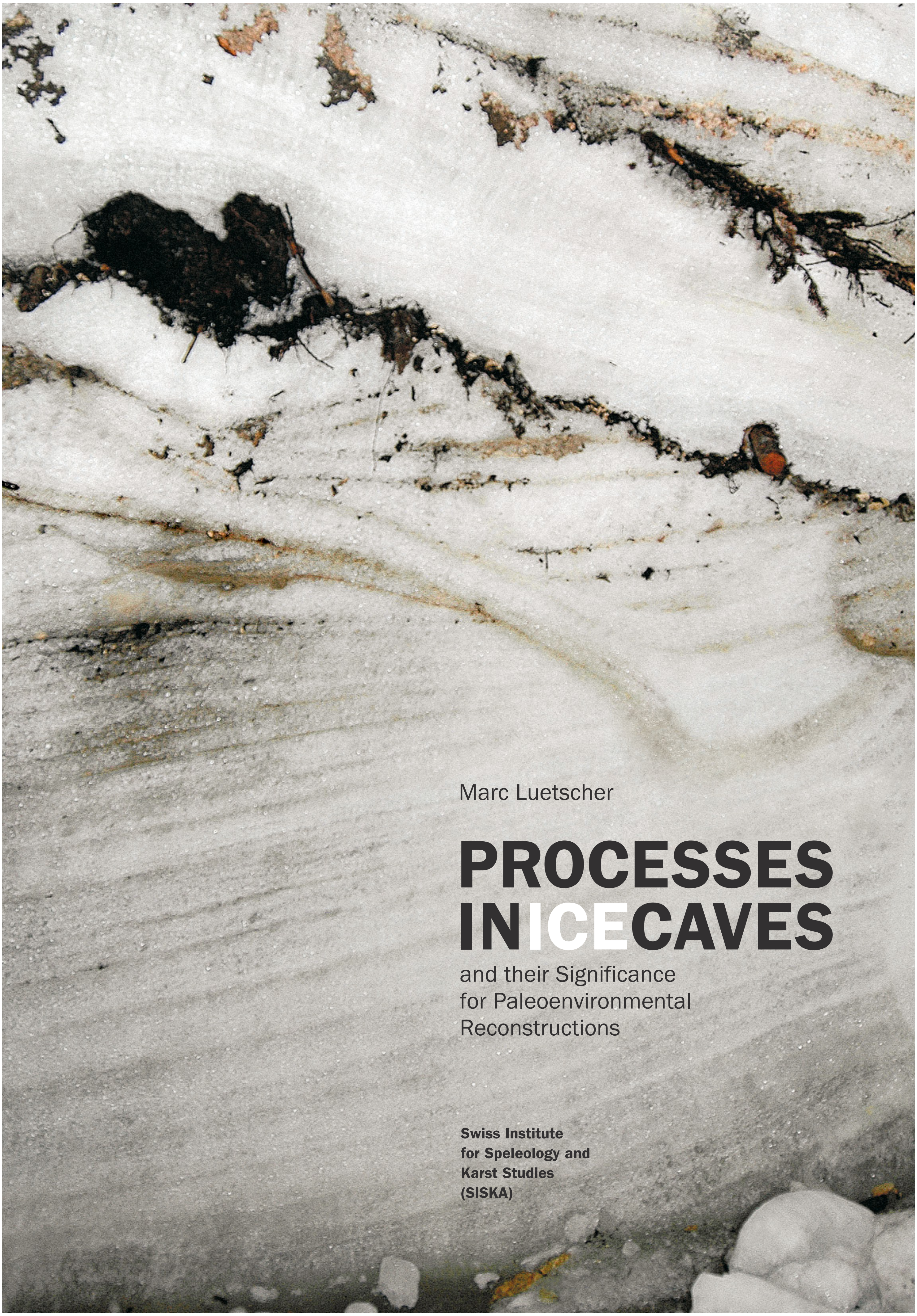

Marc Luetscher

PROCESSES IN ICE CAVES

and their Significance for Paleoenvironmental Reconstructions

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Reconstructions

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A Isaline, à Clément

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* | Foreword and acknowledgements

I was eleven when I first discovered the St-Livres ice cave in the Swiss Jura Mountains. Since then, I have been deeply fascinated by the richness found in caves. Exploring this environment with a caving club aroused my curiosity and instilled in me a desire for further scientific research. The need to understand the processes shaping our landscape prompted me to study Earth Sciences, which soon led me to discover the enormous potential of caves as a context for the understanding of past environments.

Fifteen years after my first contact with our underground realm, I had the opportunity to start a PhD on ice caves in the Jura Mountains. Such a work would not have been possible without the support, the encouragement and the contribution of numerous persons even remotely connected with the initial project.

The thesis was made possible by P.-Y. Jeannin and the Swiss speleological community as a whole. Their enthusiasm led to the creation of the Swiss Institute for Speleology and Karst Studies (SISKA) in 2000 which provided the structure for a centre of competence devoted to the study of karst, unique in Switzerland. I would like to acknowledge Pierre-Yves for the countless invaluable discussions and his trust in me since the beginning of the thesis. Even more than serving as a careful and helpful supervisor of my PhD, he was a steady motivator and supporter of my work right from the beginning.

By inviting me to join his Permafrost group, W. Haeberli provided the academic support for carrying out the thesis project. The precious contacts initiated with numerous scientists studying the Swiss cryosphere provided an new sight inside our caves. The thesis was funded by grants from the Swiss National Science Foundation (SNF 21-63764.00 and 20-103538/2).

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Many contacts with speleologists from around the world were initiated during the development of the thesis. Memorable expeditions and fieldwork were shared with Russian, Romanian, Hungarian, Austrian and Swiss speleologists. I am grateful to all of them for the enjoyable time spent together.

Several individuals actively participated in the final production of the thesis. Sue Dotson and Gina Fiore are thanked for their correction and editing work of the English writing. Christine Luetscher made an impressive final layout of the present monograph.

And last but not least, I am grateful to my family who always encouraged me in the path I chose. This thesis owes much to their continuous support and patience over these four years devoted to ice caves.

* | Summary

The effects of climate changes on the Earth ecosystem now reach extreme dimensions as compared to the Holocene timescale. In order to assess the consequences on the plant and animal kingdom, a detailed understanding of our environment is of fundamental importance. Subsurface ice accumulations provide a suitable environment for the study and documentation of past and present changes with a high temporal resolution. Located in regions where the exterior mean annual air temperature is sometimes far above 0 °C, these «ice caves» are supposed to be highly sensitive to climate changes. Thus, documenting and analysing this archive is of great interest when examining a variety of environmental parameters in space and time.
Mentioned since the sixteenth century, the presence of cave ice has been documented in numerous mid-latitude alpine karst regions. While most of the recent studies were concerned with glaciological observations, the physical understanding of the phenomenon remains incomplete. Although numerous empirical approaches identified the major processes at the origin of cave ice, their respective importance has only rarely been quantified. Therefore little data is available to predict the evolution of subsurface ice occurrences in a changing climate context.
This issue was approached from the thermodynamics perspective which considers the ice caves as distinct systems, well delimited in space. At the system boundaries, heat fluxes are exchanged with the surrounding environment. Heat transfers include: **1)** advection by forced convection; **2)** conduction through the cave walls; **3)** advection by water infiltrations; and **4)** solar radiation. Each of the terms can be quantified individually and the balance of these heat exchanges controls the ice volume within the cave. A case study validated this theoretical approach.
Because they are located at low-altitude (max. 1700 m a.s.l., MAAT ~5 °C) and are geographically well defined, the Jura mountains were considered as an ideal study site. The synthesis of the speleological literature and historical descriptions enabled recognition of nearly 25 caves containing a perennial ice accumulation. The Monlési ice cave (Boveresse, Neuchâtel, CH) was selected for a detailed investigation of its energy balance. A topographic survey of the cave assessed the ice volume at nearly 6000 m^3 for an ice thickness of about 12 m. Field observations showed a filling mainly constituted of congelation ice and partly fed by the firn accumulated at the base of the entrance shafts. Dating attempts based on a multi-parametric approach involving the study of the ice stratigraphy, the clast typology and isotopic analyses (^3H, ^{210}Pb, ^{14}C) assessed the age of the accessible ice volume at about 120 years. This value implies an energy supply of about $1 Wm^{-2}$ at the ice-rock interface which was validated by the melting rates observed at the base of the ice volume between 2001 and 2004 (~10 cm $year^{-1}$) and by temperature measurements performed in four 80 cm deep boreholes located in the lateral cave walls.
Although snow accumulations partially compensate for this heat supply, the cooling of the ice cave is mostly attributed to the strong air circulation which can be observed within the cave. During the winter season, when the external temperature decreases below the cave air temperature, a forced convection is initiated between the three entrance shafts. Measurements performed at Monlési ice cave have shown that air fluxes could reach over 10 m^3/s. The sublimation associated with this natural ventilation represents a major factor in the heat exchange between the cave and the surrounding environment. Conversely, during the summer, when the external temperature increases above the subsurface air temperature, the cave acts as a cold air trap leading to a strong thermal anomaly.
The analysis of the energy balance of Monlési ice cave demonstrates that ice volume fluctuations reflect the imbalance observed between the energy loss («cooling») and the energy supply («warming») of the cave. Although many uncertainties regarding each factor involved in the energy balance remain, the relative importance of the different parameters was assessed. The quantification of heat exchanges at the Glacière de Monlési clearly demonstrated that the cave ice mass balance is closely related to: **1)** the winter cooling induced by air circulation (forced convection), which is responsible for 70% of the heat loss; and **2)** the latent heat associated with winter snow accumulations at the base of the entrance shafts, which is responsible for about 20% of the heat loss.
The process-based model of Monlési ice cave suggests that the heat exchange induced by the forced convection could be significantly reduced in a warming climate context (less 30% if the MAAT increases of 1 °C). However, the efficiency of heat transfer increases if the freezing of percolating meltwater enables the formation of cave ice. This is possible only in the presence of sufficient amounts of water. But, since the crystallisation becomes almost impossible with elevated discharge rates (i.e. major precipitation events), the optimal equilibrium state is reached with daily melting of the exterior snow cover followed by freezing nights. This condition is preferentially reached during the early spring, but mid-winter spells also represent a favourable context for the formation of large amounts of cave ice. Since the frequency of such oscillating freeze/thaw cycles increases significantly with a warming climate, an unexpected positive cave ice mass balance may be observed if external temperatures are fluctuating around 0 °C. Hence, our study suggests that contrary to the Northern Hemisphere snow cover, which is projected to decrease further during the twenty-first century, some mid-latitude/low-altitude ice caves could present a stable

decadal mass balance under favourable climatic conditions. Nevertheless, due to a concurrent reduction of low-altitude snow precipitation and number of annual freezing days at low altitudes, a noticeable negative cave ice mass balance has been observed in many caves of the Jura mountains since the end of the 1980s.
The presence of cave ice is controlled by specific winter climatic conditions (i.e. winter precipitation regime, number of freezing days, mean annual air temperature), therefore, ice caves can be considered as valuable environmental markers. It was demonstrated that cave ice can be preserved over several centuries if not thousands of years (for instance, Glacière de St-Livres: 1200 y. B.P.). Located close to the source regions of atmospheric pollutions, mid-latitude/low-altitude cave ice represents an important archive of climate history in Central Europe. Therefore, further investigations should focus on the interpretation of high resolution glaciochemical analyses of cave ice cores. A numerical model of evaporative processes along a karst conduit would be a useful tool for the interpretation of the stable isotope records.
This study has shown that evidence of periglacial contexts is found in several caves of the Jura Mountains where the cave ice completely melted. Because such patterns might be observed over thousands of years, our observations sustain that cave sediments could be a good marker of Holocene winter climate fluctuations.

* | Résumé

Les changements climatiques observés au cours des deux derniers siècles atteignent des dimensions exceptionnelles à l'échelle de l'Holocène. Afin d'en évaluer précisément les conséquences sur l'écosystème terrestre, une connaissance détaillée de notre environnement s'avère indispensable. Les accumulations de glace souterraine constituent des objets potentiellement favorables à l'étude et à la documentation de changements passés et présents, avec une haute résolution. Situées dans des régions où la température moyenne annuelle de l'air extérieur est parfois bien supérieure à 0 °C, ces «glacières» sont fortement exposées aux fluctuations climatiques. La documentation et l'analyse de cette archive, susceptible d'enregistrer de nombreux paramètres environnementaux, relève dès lors d'un grand intérêt paléoclimatique.
Mentionnée dès le 16ème siècle, la présence de glace souterraine est documentée dans la plupart des karsts alpins de moyenne latitude. Alors qu'une large majorité des études récentes se concentre sur les enregistrements glaciochimiques, une interprétation physique du phénomène reste lacunaire. Bien que de nombreuses approches empiriques aient identifié les principaux processus à l'origine de la glace souterraine, leur importance respective n'a, à ce jour, que rarement été quantifiée. Ainsi, peu d'éléments permettent d'anticiper l'évolution de ces glacières dans une perspective de changement climatique.
La problématique est abordée sous l'angle de la thermodynamique. Cette approche considère les glacières comme des systèmes bien délimités dans l'espace, aux limites desquels des flux de chaleurs sont échangés avec l'environnement. Ces transferts d'énergie incluent: **1)** l'advection par convection forcée; **2)** la conduction au travers des parois; **3)** l'advection par infiltrations d'eau; et **4)** le rayonnement solaire. Chaque terme pouvant être quantifié individuellement, le bilan de ces échanges de chaleur définit l'équilibre du volume de glace au sein de la cavité. Cette approche théorique est validée par une étude de cas sur le terrain.
La chaîne jurassienne a été sélectionnée comme terrain d'étude idéal. Situé à basse altitude (alt. max. 1700 m) et bien délimité géographiquement, ce site présente l'avantage d'une importante documentation spéléologique. Une synthèse de cette documentation permet de recenser quelque 25 cavités contenant une accumulation de glace pérenne. Toutefois, l'analyse des descriptions antérieures suggère que de nombreuses glacières présentent un bilan de masse négatif fortement marqué depuis la fin des années 1980. La glacière de Monlési (Boveresse, Neuchâtel, CH) a été choisie pour l'étude détaillée de son bilan énergétique. Le levé topographique de la cavité permet d'estimer le volume de glace à quelque 6000 m^3, pour environ 12 m d'épaisseur. Les observations montrent que ce remplissage

est essentiellement constitué de glace de congélation, mais partiellement alimenté par les névés formés à la base des puits d'entrée. Les essais de datation basés sur une approche multiparamétrique incluant l'étude de la stratigraphie de la glace, de la typologie des clastes ainsi que des analyses isotopiques (^{3}H, ^{210}Pb, ^{14}C) permettent d'estimer l'âge maximal du volume de glace accessible à quelque 120 ans. Cette valeur implique un apport de chaleur d'environ $1 Wm^{-2}$ à l'interface roche-glace, validé par la fonte observée à la base du volume de glace entre 2001 et 2004 (~10 cm an^{-1}) et par les mesures de températures menées dans 4 forages de 80 cm de profondeur répartis dans les parois latérales de la cavité.

Les accumulations de neige hivernale permettent de partiellement compenser cet apport de chaleur. Néanmoins, le refroidissement de la glacière est principalement attribué aux importantes circulations d'air observées au sein de la cavité en période hivernale. Lorsque la température extérieure décroît en deçà de la température hypogée, une convection forcée s'établit entre les trois puits d'entrée de la cavité. Les mesures effectuées à la Glacière de Monlési montrent des flux d'air pouvant atteindre plus de 10 m^3/s. La sublimation associée à cette ventilation naturelle constitue dès lors un élément majeur de l'échange thermique affectant la cavité. Au contraire, en été, lorsque la température extérieure dépasse la température hypogée, la cavité agit comme piège thermique induisant une forte anomalie thermique au sein de la cavité.

L'analyse du bilan énergétique de la glacière démontre que les fluctuations du volume de glace reflètent le déséquilibre observé entre la perte («refroidissement») et l'apport («réchauffement») d'énergie. Bien que de nombreuses incertitudes demeurent sur chacun des termes du bilan énergétique, l'importance relative des différents paramètres est démontrée. La quantification des échanges de chaleur à la Glacière de Monlési démontre clairement que le bilan de masse de la glace souterraine est tributaire de: **1)** un refroidissement hivernal induit par les circulations d'air (convection forcée) responsables de près de 70% des pertes de chaleur; et **2)** la chaleur latente liée aux accumulations de neige hivernale à la base des puits d'entrée.

La formation de glace n'est cependant possible qu'en présence de quantités suffisantes d'eau. Dès lors, les conditions les plus favorables à la cristallisation de glace s'observent en période hivernale, lorsque la température extérieure est supérieure à 0 °C (permettant la percolation d'eau de fonte ou de pluie), alors que la température hypogée demeure en deçà du point de congélation. Ces conditions sont préférentiellement atteintes au début du printemps, lorsque les nuits glaciales succèdent aux journées plus chaudes. Néanmoins, des périodes de redoux constituent également un contexte favorable à la cristallisation d'importantes quantités de glace.

La présence de glace souterraine étant contrôlée par des conditions climatiques particulières (i.e. précipitations hivernales, jours de gel, température moyenne annuelle), les glacières peuvent être considérées comme des marqueurs environnementaux fiables. Le modèle physique de la Glacière de Monlési suggère que, dans un contexte de réchauffement climatique, une réduction significative de la convection forcée doive être escomptée. Toutefois, l'efficacité des transferts de chaleur croît en présence de percolations d'eau de fonte. L'équilibre optimal est atteint lors de fontes journalières suivies par des nuits glaciales. Etant donné qu'une telle oscillation de cycle de gel/dégel augmente significativement avec un réchauffement climatique, un bilan de masse positif de la Glacière peut être observé si les températures de l'air extérieur fluctuent autour de 0 °C. Ainsi, contrairement à la couverture neigeuse de l'hémisphère nord, certaines glacières de moyenne-latitude/basse-altitude pourraient présenter des bilans de masse décennal équilibrés en présence de conditions climatiques favorables.

Il a été démontré que cette glace souterraine pouvait être conservée durant plusieurs milliers d'années. Située à proximité des régions sources de pollutions atmosphériques, la glace de moyenne-latitude/basse altitude constitue une importante archive de l'histoire climatique de l'Europe centrale. Les études futures devraient dès lors se concentrer sur l'analyse chimique détaillée de carottes complètes issues des glacières. Un modèle numérique des processus d'évaporation le long d'un conduit karstique constituerait pour cela un outil précieux en vue de l'interprétation des données.

En outre notre étude a permis de documenter des signatures périglaciaires dans de nombreuses cavités où la glace a complètement disparu. Etant donné que ces signatures peuvent être conservées durant plusieurs milliers d'années, nos données soutiennent que les sédiments de grottes peuvent constituer de bons marqueurs pour l'étude des fluctuations du climat hivernal durant l'Holocène.

* | Zusammenfassung

Die Effekte der Klimaänderungen auf das terrestrische Ökosystem erreichen heute extreme Masse im Vergleich zu den Ereignissen auf der Holozänskala. Um die Konsequenzen einzuschätzen ist ein detailliertes Verständnis unserer Umwelt von grundlegendem Wert. Unterirdische Eisablagerungen stellen für die Untersuchung von Klimaveränderungen ein gut verwendbares Archiv dar.

Eishöhlen sind auf Klimaänderungen hoch empfindlich, da sie in Regionen vorkommen, wo die durchschnittliche Aussentemperatur manchmal weit über 0 °C steigt. Es ist aufschlussreich, dieses Archiv zu dokumentieren und zu analysieren, da es eine Vielzahl von räumlichen und zeitlichen Klimaparametern anbieten könnte.

Höhleneis ist seit dem sechzehnten Jahrhundert bekannt und wurde in zahlreichen Karstregionen niederer Breite dokumentiert. Die meisten aktuellen Studien beschäftigen sich jedoch mit glaziologischen Beobachtungen, so dass das physikalische Verständnis des Phänomens unvollständig ist. Obgleich zahlreiche empirische Untersuchungen die Hauptprozesse am Ursprung des Höhleneises beschreiben, wurden diese nur selten quantifiziert. Folglich sind nur wenige Daten vorhanden um die Entwicklung des unterirdischen Permafrostes in einem ändernden Klimakontext vorauszusagen.

Diese Studie wird unter dem Aspekt der Thermodynamik behandelt. Demnach sind Eishöhlen als räumlich gut abgegrenzte Systeme anzuschauen, an deren Grenzen Wärmeflüsse mit der Umwelt ausgetauscht werden.

Diese beinhalten: **1)** Advektion durch Zwangskonvektion, **2)** Übertragung durch die Höhlenwände, **3)** Advektion durch Wasserinfiltration und **4)** Sonneneinstrahlung. Jeder Wärmefluss kann einzeln quantitativ bestimmt werden, und die Bilanz dieser Wärmeflüsse bestimmt das Eisvolumen innerhalb der Höhle. Eine Fallstudie validiert diesen theoretischen Ansatz.

Weil der Jura auf niederer Höhe liegt (max. 1700 m ü.M.) und geografisch gut definiert ist, wurde diese Bergkette als ideal für eine solche Studie betrachtet. Die Synthese der speläologischen Literatur und der historischen Beschreibungen ermöglichte die Identifizierung von 25 Höhlen, die dauerhafte Eisansammlungen enthalten. Monlési Eishöhle (bei Boveresse, Neuchâtel, CH) wurde für eine ausführliche Untersuchung der Energiebilanz ausgewählt. Die Kartierung ergab ein Eisvolumen von ca. 6000 m^3 und eine Eisstärke von ungefähr 12 m. Beobachtungen zeigten, dass die Füllung hauptsächlich durch Gefriereis entsteht, teilweise auch durch den Firn, der sich an der Basis der Eingangsschächte ansammelt. Die Eisdatierung geht von einem multi-parametrischen Ansatz aus, der die Eisstratigrafie, die Typlogie der Clasten und isotopische Analysen einschliesst (^{3}H, ^{210}Pb, ^{14}C). Mit dieser Methode wurde das Alter des zugänglichen Eisvolumens auf etwa 120 Jahren geschätzt. Aus diesem Wert resultiert ein Energieaustausch an der Eis-Felsen Schnittstelle von ungefähr 1Wm^{-2}, der durch die beobachtete Schmelzrate an der Eisbasis zwischen 2001 und 2004 (~10 Zentimeter pro Jahr) sowie durch Temperaturmessungen, durchgeführt in vier 80 Zentimeter tiefen Bohrlöchern in den seitlichen Höhlenwänden, validiert wurde. Obwohl Schneeablagerungen diesen Energiefluss teilweise ausgleichen, wird das Abkühlen der Eishöhle hauptsächlich der im Innern der Höhle beobachteten, starken Luftzirkulation zugeschrieben. Während der Winterzeit sinkt die Lufttemperatur ausserhalb unter die Lufttemperatur in der Höhle, so dass eine Zwangskonvektion zwischen den drei Eingangsschächten eingeleitet wird. Die Messungen in der Monlési Eishöhle haben gezeigt, dass die Luftflüsse Werte von über 10 m^3/s erreichen können. Die Sublimation, die mit dieser natürlichen Ventilation verbunden ist, spielt eine wichtige Rolle im Wärmeaustausch zwischen der Höhle und ihrer Umgebung. Während des Sommers, wenn die Aussenlufttemperatur über die Höhlenlufttemperatur steigt, wird die Höhle zur Kältefalle, die zu einer starken thermischen Abweichung führt.

Die Analyse der Energiebilanz der Monlési Eishöhle zeigt, dass die Eisvolumenfluktuationen die Unausgeglichenheit zwischen dem Energieverlust («Abkühlung») und dem Energiezusatz («Aufwärmung») der Höhle reflektieren.

Obwohl zu jedem Faktor der Energiebilanz weiterhin Ungewissheiten bestehen bleiben, kann der relative Wert der verschiedenen Parameter festgesetzt werden. Die Quantifikation der Wärmeaustausche in der Monlési Eishöhle zeigt, dass die Massenbilanz des Höhleneises eng zusammenhängt mit **1)** der Winterabkühlung durch Luftzirkulation (Zwangskonvektion), die für 70% des Wärmeverlust verantwortlich ist, und **2)** der latenten Wärme der Winterschneeablagerungen, die für ungefähr 20% des Wärme-verlust verantwortlich ist.

Das Prozess-basierte Modell der Monlési Eishöhle sagt voraus, dass der Wärmeaustausch, der durch die Zwangskonvektion verursacht wird, in einem zunehmend wärmeren Klimakontext erheblich verringert würde (30% weniger bei einer Zunahme von 1 °C der jährlichen Durchschnittstemperatur). Jedoch erhöht sich die Effizienz der Wärmeübertragung, wenn das Einfrieren des sickernden Schmelzwassers die Bildung von Höhleneis ermöglicht. Dies ist angängig davon, dass genügend Wasser vorhanden ist. Da die Kristallisation mit erhöhten Abflussraten (d.h. grösseren Niederschlägen) fast unmöglich wird, ist der optimale Gleichgewichtszustand mit dem täglichen Schmelzen der Aussenschneeabdeckung und dem nächtlichen Einfrieren erreicht. Typischerweise ist dies während des frühen Frühlings der Fall, Schmelzereignisse mitten im Winter sind jedoch auch vorteilhaft für die Bildung von grossen Mengen Höhleneis. Da die Frequenz solch oszillierender Kristallisation/Schmelz-Zyklen mit einem sich erwärmenden Klima

erheblich zunehmen, könnte eine unerwartete positive Höhleneis Massenbilanz beobachtet werden, wenn die externen Temperaturen um 0°C schwanken. Folglich kann davon ausgegangen werden, dass im Gegensatz zur Schneeabdeckung der Nordhemisphäre, die sich während des einundzwanzigsten Jahrhundert wahrscheinlich verringern wird, gewisse tief gelegene Eishöhlen unter veränderten klimatischen Verhältnissen ausgeglichene zehnjährige Eismassenbilanzen aufweisen werden. Dennoch wird in vielen Höhlen des Juras seit dem Ende der achtziger Jahre wegen der gleichzeitigen Reduktion der Schneeniederschläge und der Zahl der Frosttage auf niederen Höhen eine wahrnehmbare negative Höhleneismassenbilanz beobachtet.

Da Höhleneis von spezifischen Winterverhältnissen gesteuert wird (d.h. Winterniederschlagsregime, Zahl der Frosttage, jährliches Mittel der Lufttemperatur) und da dies in der Nähe der Quellregionen der atmosphärischen Verunreinigungen liegt, können Eishöhlen als wertvolle Klimaarchive betrachtet werden. Es wurde gezeigt, dass sich das Höhleneis über mehrere Jahrhunderten konservieren kann, wenn nicht tausende von Jahren (so z.B. der Glacière de St-Livres: y. 1200 BP). Folglich sollten sich weitere Untersuchungen auf die Interpretation von hoch aufgelösten glaziochemischen Analysen der Höhleneiskerne konzentrieren. Ein numerisches Modell von Verdampfungsprozessen entlang einer Karströhre würde dabei ein nützliches Werkzeug für die Interpretation von Isotopenaufzeichnungen sein.

Die Studie beweist periglaziale Kontexte auch in Höhlen des Juras, in denen das Höhleneis vollständig geschmolzen ist Weil solche Vorkommen über tausende von Jahren beobachtet werden können, betonen unsere Beobachtungen nahe, dass Höhlensedimente als gute Archive der holozänen Klimaschwankungen betrachtet werden können.

PART I:

PROCESSES IN ICE CAVES

and their Significance
for Paleoenvironmental
Reconstructions

a case study
from the Jura Mountains

1 | Introduction

The effects of climate changes on the Earth ecosystem now reach extreme dimensions as compared to the Holocene timescale (~last 10'000 years). In order to assess the consequences on the plant and animal kingdom (including the socio-economical aspects), a detailed understanding of present and past environmental changes is of fundamental importance (cf. Oldfield & Alverson, 2003). Much of our knowledge about variations in past climate has been gained over the last 40 years through the analysis of isotopic composition of marine sediments. More recently, data issued from ice cores performed in Greenland (Dansgaard et al., 1993) and Antarctica (Petit et al., 1999) provided complementary records of global climate changes. These data are rapidly considered as a universal template for paleoclimatic research (Oldfield, 2003). But since it reflects only poorly the climatic fluctuations observed at a regional scale, the interest in mid-latitude continental records has grown significantly in recent years. Examples of such continental records are lake sediments, peat bogs, cave sediments and glaciers.

The cryosphere, particularly exposed to climatic changes, constitutes a study object of special interest. Climate projections suggest a drastic decrease of the glacial cover during the twenty-first century (e.g. IPCC, 2001) and the lower limits of permafrost occurrences could rise by several hundred meters in the coming decades. In cold mountain areas, such changes increase the risk of natural hazards with respect to climate changes. Haeberli and Burn (2002) identified among these hazards the retreat of glaciers, rock falls from destabilized mountain walls, ice avalanches, mudflows from glacier-clad volcanoes, floods from ice- and moraine-dammed lakes and debris flows from steep slopes or breaching moraines. The related fluctuations of the spatial distribution of glaciers and permafrost outline regional modifications of the energy balance at the Earth's surface. Therefore, in mountain areas, shifts in glacial cover can be considered as reliable key indicators for the early detection of present-day climate changes (e.g. Hoelzle et al., 2003; Nesje & Dahl, 2003). Located close to the source regions of atmospheric pollutions, glacial ice constitutes an important archive of climate history in Central Europe. In the alpine belt, suitable sites for paleoclimatic studies were identified (Funk, 1994) and ice coring was performed successfully at four selected sites: Colle Gniffeti and Upper Grenzgletscher, Monte Rosa, Switzerland (Oeschger et al., 1977; Schotterer et al., 1985; Wagenbach et al., 1988; Döscher et al., 1995; Schwikowski et al., 1999a; Eichler et al., 2000; Luethi & Funk, 2001); Fiescherhorn, Aar Massif, Switzerland (Schwikowski et al., 1999b); and Col du Dôme, Mt-Blanc Massif, France (De Angelis & Gaudichet, 1991; Vincent et al., 1997; Preunkert et al., 2000). Glacio-chemical investigations performed on these ice cores enabled translation of proxy data into paleoclimatic and paleoatmospheric records by comparing them with corresponding historical records and data from meteorological and air quality measurements (Schwikowski, 2004). Ice archives from pre-industrial times also provide information about natural concentration levels for various atmospheric species which are now influenced by human activities (Harnisch et al., 1996). Therefore, paleo-records from alpine glaciers and miniature ice caps allow the reconstruction of European atmospheric pollution history (e.g. Schwikowski, 2004). Although flow-modelling suggests that the basal ice in the Monte Rosa (Swiss Alps) area might range between 1000 and 2000 years old (Schotterer et al., 1985; Luethi & Funk, 2001), ice cores performed to date have only recovered records that are between 30 and 200 years old. Hence, an interest in longer glacial records remains real.

The study of subsurface ice is of evident paleoclimatic interest because data suggests the presence of subsurface ice that is more than 3500 years old (e.g. Pop & Ciobanu, 1950; Lauriol & Clark, 1993; Schroeder, 1977). Although they were described for centuries, studies of such «ice caves» have been limited (e.g. Hill & Forti, 1997). Most of these investigations were concerned with morphological observations (e.g. Kyrle, 1923; Racovita et al., 1987; Maire, 1990) but some studies focused on the climatic processes at the origin of cave ice (e.g. Bock, 1913; Saar, 1956; Racovita, 1972). However, empirical data collected during the twentieth century must now be completed by robust process-based models which can be validated by field observations subject to fast environmental changes.

Mid-latitude/low-altitude ice caves provide a suitable environment for the study and documentation of past and present changes with a high resolution. Located in regions where the exterior mean annual air temperature is sometimes far above 0 °C, subsurface ice patterns are supposed to be highly sensitive to climate changes. In a context of climate warming, it appears to be of fundamental interest to document and analyse this archive assumed to offer a variety of environmental parameters in space and time.

The preliminary identification of exterior climate parameters which control processes in ice caves is indispensable for a paleo-environmental interpretation. Therefore, the present work focuses on the climatic study of ice caves. The physically-based, process-oriented model which emanates from this study relies on the energy balance at a selected study site. The validation of the model is provided by the quantification of heat exchanges determined from field measurements. Although major approximations had to be made on processes which are not explicitly included in the model (for instance, exterior meteorological conditions, hydrological conditions at the surface), the heat balance became a valuable tool in identifying the main processes at the origin of an ice cave. Therefore, our model enables observations performed on variations of cave ice mass balances to be placed into a broader context of climatic change.

The present study demonstrates the major role of winter climatic conditions in the conservation of mid-latitude/low-altitude ice caves, while summer climate has little influence on the cave ice mass balance. However, since several morphological parameters are considered in the energy balance (including the dimension of the entrance shafts and the thermal exchange surface), the model also underlines the differences between the observed ice caves. These differences are particularly well set in evidence by the mass turnover rate of the cave ice. Datings performed in two ice caves of the Jura Mountains help distinguish a very active system showing mass turnover of ~120 years (Monlési, Neuchâtel) from a much less active system showing an age of over 1200 years (St-Livres, Vaud).

The results presented in this manuscript are organized in two parts. The first part starts with a historical background of studies related to ice caves. Besides providing a scientific context, this approach enables the proposal of a synthetic conceptual model of the processes at the origin of cave ice. This step is necessary to raise the most pertinent questions for designing the instrumentation of the studied ice caves. Field measurements are summarized within the chapter about the main results of the study. A broad discussion of the results follows, which interprets the major findings according to the present understanding of ice caves. The present study provides a first complete and validated process-based model of ice caves. The second part of the manuscript comprises a number of publications related to the study. These papers give more details on data, models and interpretations than summarized in the first part. Although raw data are not integrated into this manuscript, they can be consulted online at www.isska.ch. A selected bibliography of ice cave related literature completes the manuscript. A comprehensive bibliographical list is available at the ISSKA website.

2 | Scientific background

2.1 Historical developments

Low-altitude subsurface ice occurences in the Jura Mountains were first mentioned in manuscripts of the sixteenth century (e.g. AEN, 1554; Poissenot, 1586), but archaeological remains suggest that at least some of these ice caves were known during the Romanian epoch (Girardot & Trouillet, 1885). In fact, because of the proximity of nearby human settlements, cave ice was frequently exploited for personal use or regional trade by shepherds or local inhabitants (e.g. Browne, 1865; Gauchon, 1997). Easily accessible and well known by native people, ice caves of the Jura Mountains (France and Switzerland) have been the object of numerous descriptions for hundreds of years (e.g. Boisot, 1686; Billerez, 1712). Recognising ice caves as a particular phenomenon, the interest of naturalists has grown during the eighteenth and nineteenth centuries (Fig. 1). Hence, temperature measurements and meticulous observations were soon carried out in different caves (for instance De Cossigny, 1750 [in Deluc, 1822]; Prévost, 1789). Early hypotheses on ice formation were sometimes the object of vigorous discussions (e.g. Deluc, 1822), and the first consistent scientific explanations for the presence of cave ice based on field observations are attributed to Thury (1861). This author first suggested the major role of cold air circulation in the presence and the conservation of subsurface ice fillings.

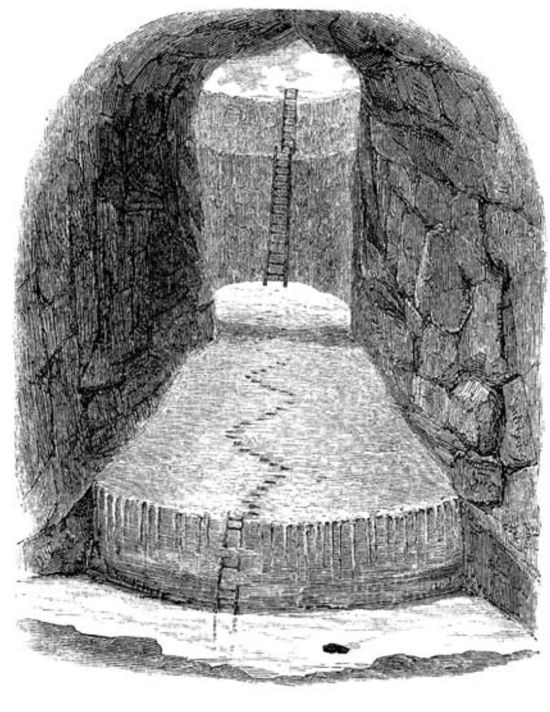


Fig. 1 Engraving of the Glacière de St-Livres, Switzerland (Browne, 1865)

However, the interest in ice caves was not limited to the Jura Mountains; such features were recognised in other parts of Europe (e.g. Steno, 1671) and led to numerous descriptions in the regional literature. New observations and studies were soon initiated in Austria (e.g. Braune, 1802 [in Mais, 1999]), Romania (e.g. Schmidl, 1863 [in Racovita & Onac, 2000]) and Slovakia (e.g. Ruffiny, 1870 [in Droppa, 1960]). The abundant literature rapidly becoming available encouraged various authors to propose larger reviews of observations carried out on such caves (e.g. Girardot & Trouillet, 1885; Fugger, 1888; Balch, 1900).

2.2 Worldwide literature related to ice caves

Early observations of caves were generally reported in the form of short descriptions within books focused on a variety of topics completely unrelated to caves (for instance philosophy or poetry). Consequently, citations of ice caves can be found in several diaries of journeys in the alpine belt (e.g. Steno, 1671; De Saussure, 1779). Although publications with an increasing scientific interest became more frequent, papers related to ice caves remained scattered through non-specific periodical journals until the end of nineteenth century. In fact, many early observations were presented to European natural science societies (for instance, *Mémoire de l'Académie*, *Mitteilungen der Gesellschaft für Salzburger Landeskunde, Mémoires de la Société d'Emulation du Doubs, Annales de Chimie et Physique,* etc.). After the first caving journals appeared in 1895, descriptions and observations completed by speleologists on ice caves were published in one of the numerous regional speleological periodicals. According to Forti (2004), there were at least 2000 caving journals worldwide by the end of the twentieth century. The extremely high number of publications, with an average of very low print numbers and irregular issues (if not sudden death), makes it extremely difficult to search for articles on a given caving topic. Fortunately, since 1961, the annual publication of Speleological Abstracts by the Swiss Society of Speleology, supported by the International Union of Speleology, facilitates easier access to recent international cave literature references. To foster information exchange within the «ice cave community», a list of bibliographical references dedicated to ice caves was compiled (cf. www.isska.ch). The *Bibliography of Ice Caves* is a searchable database, compiled by collaborating researchers involved in the study of ice caves. Accessible by many scientists, the database contains over 400 citations of works published worldwide since 1554.

2.3 Major study sites

During the twentieth century, some of the documented ice caves became the object of larger studies. Among the three most investigated sites, the Ghetarul de la Scarisoara (RO), was probably the first to be described in detail by E. Racovitza (1927 [in Serban et al., 1967]), and it certainly remains one of the most studied ice caves (e.g. Racovita & Onac, 2000). Located in the Romanian Carpathians at 1165 m above sea level (a.s.l.), this cave is composed of a single entry shaft of 60 m in diameter and 48 m in depth. Its morphology contributes to a large snow accumulation during winter and acts as an important cold air trap in summer (Racovita, 1975).
In the Austrian Alps, the Rieseneishöhle in the Dachstein region was studied extensively over the last 100 years (e.g. Mais, 1999; Pavuza & Mais, 1999). With a lower entrance located at about 1500 m a.s.l., major air circulation impacts the freezing of the ice in this cave. Furthermore, the Eisriesenwelt (Tennengebirge, AU), commonly considered as the largest ice cave in the world, was the object of several investigations after its discovery (Kyrle, 1923; Angermayer et al., 1926) but no studies have been conducted in this cave in recent decades. Discovered in 1870 in Slovakia, Dobsinska Ice Cave was opened to tourist exploitation one year later. Climatic measurements were made at the end of nineteenth century and more recent studies were conducted over the past twenty years, including geophysical investigations (Lalkovic, 1995). Throughout the world, several other ice caves were investigated (e.g. Bini & Pellegrini, 1998; Dublyansky & Kadebskaya, 2003; Ohata et al., 1994 a & b). Despite an initial rush in studying these caves in the Jura Mountains, the only ice cave in this massif which was investigated further is the Glacière de Monlési, where detailed descriptions were documented (Stettler & Monard, 1960; Stettler, 1971; Pancza, 1992). In the early years, methods applied to the study of ice caves were essentially focused on speleological observations (cave mapping, description, observations). However, improvements in data acquisition technology facilitated short-term climatological investigations, and monitoring became more frequent in recent decades (e.g. Perroux, 2001). Hence, the understanding of processes at the origin of ice caves has increased significantly.

2.4 State of the art

2.4.1 Ice cave distribution

Considering the distribution of ice caves as a function of external freezing and thawing indices, Harris (1979; 1982) observed that the distribution of Canadian ice caves correlates well with common boundaries of permafrost zones. In temperate latitudes, spatial modelling of permafrost distribution patterns as a function of Mean Annual Air Temperature (MAAT) suggested that permafrost may exist at altitudes as low as 500 m a.s.l. (i.e. MAAT ~10 °C) where there is no solar radiation (Hoelzle et al., 1993). Although the presence of low-altitude cave ice has been recognized in many different regions (e.g. Maire, 1990), no recent inventory of its distribution has been performed.

2.4.2 Glaciological investigations

Cave ice morphology was investigated by many authors who meticulously described numerous crystallization forms (e.g. Racovita, 1994; Schroeder, 1977; Dorofeeva, 1989). The crystallization forms were attributed to three main origins: diagenesis of snow accumulations (firn), freezing of infiltration water (congelation ice) and freezing of condensation water (hoarfrost) (see also Schumskii, 1964 for further details).

Based on observations carried out in Ghetarul de la Scarisoara, Racovitza (1927 [in Viehmann, 1973]) distinguished seasonal ice from fossil ice, which seems to not be affected by seasonal variations. Likewise, by observing the absence of significant fluctuations of the cave ice mass balance, Marshall (1975 [in Schroeder, 1977]) suggested the presence of relict cave ice in the Canadian Rocky Mountains.

However, significant ice movements are initiated by melting at the rock-ice interface and were occasionally well documented (Tulis, 1996; Serban & Racovita, 1991). In Ghetarul de la Scarisoara, Silvestru (1999) estimated this basal melting at 6.5 mm year^{-1}, which was chiefly attributed to lithological pressures. As suggested by Brown & Marshall (1973), a description of the cave ice dynamics enables a better estimation of melting/accretion rates. But, next to seasonal variations, ice extensions indicate multi-annual oscillations of increasing or decreasing ice volume (e.g. Serban & Racovita, 1991). In Ghetarul de la Scarisoara, Racovita et al. (1987) observed that, between 1960 and 1989, at least 35 cm of the upper layers of the ice block melted. Noticing a global ice decrease of more than 147 cm between 1947 and 1973 in this cave, Viehmann (1973) estimated that an increase or preservation of the ice is no longer possible under the present climatological conditions in Europe.

A gradual disappearance of low-altitude cave ice is supported by numerous observations in Europe and overseas (e.g. Sesiano, 1996; Ohata et al., 1994b; Trimmel, 1969). Pavuza & Mais (1999) mentioned an ice reduction of about 2 cm year^{-1} since the beginning of measurements in 1995 in Dachstein-Rieseneishöhle. Next to global climatic variations (e.g. Saar, 1956), qualitative appreciations are given on the influence of various parameters of ice formation or conservation. Trimmel (1969), using the example of Beilstein-Eishöhle, Austria, attributes the presence of low-altitude ice caves to the dense vegetation coverage. He suggests that the lower effective infiltration, resulting from evapotranspiration processes taking place at the surface, enables a better conservation of the ice in comparison to uncovered karst regions. Vincent (1974) enhanced the importance of cave structure and winter cold for the presence of subsurface ice but secondary factors such as cave location, surface vegetation and availability of moisture are mentioned for ice caves in the Pryor Mountains, US.

As long-term observations remain sporadic, a few authors tried to demonstrate ice fluctuations by comparing recent observations to old pictures and/or written descriptions (Brulhart, 2001; Pavuza & Mais, 1999) or maps (Luetscher & Wenger, 2002). Therefore, only few quantified mass balances are available.

2.4.3 Assessment of cave ice volume

Geophysical investigations were carried out essentially to determine the total ice thickness in places where direct observations were almost impossible or to confirm the estimated values. Owing to the frequent presence of cryoclastic sediments (e.g. Pancza, 1992), the use of geophysical methods is complicated. Furthermore, common investigation methods need special adaptations to avoid scattering which is induced by the three-dimensional characteristics of the cave (for instance, ground penetrating radar-GPR). Nevertheless, GPR surveys were successfully performed in Dobsinska ice cave (Geczy & Kucharovic, 1995; Novotny & Tulis, 1995) to determine the ice thickness.

Ultrasonic probes (0.5 MHz) were conducted in Scarisoara ice cave (Romania), and provided a detailed picture of the structure of the upper ice layers (Silvestru & Boghean, 1992). Unfortunately, they could not provide an investigation depth of more than five meters. Although no deep drillings were done, the ice filling of Scaricoara was estimated to be about 24 m thick, which represents more than 1000 ice layers of 0.5 - 15 cm (Silvestru & Boghean, 1992).

Measuring these thicknesses allows a better estimation of the ice volumes, which can vary from only a few cubic meters to more than 100'000 m^3 (the ice filling in Dobsinska was estimated at 110'132 m^3 (!) by Novotny & Tulis, 1996). In Ghetarul de la Scarisoara, the ice filling is considered to be between 50'000 to 75'000 m^3 (Serban et al., 1948 [in Serban et al., 1967]; Silvestru, 1989).

However, because precise measurements are rarely possible, most authors based their estimations on direct field observations. Thus, the related uncertainty is probably about 30% of the estimated value.

2.4.4 Ice dating

In several caves, the presence of «rimayes» (crevasses or gaps between ice and bedrock) often provides an excellent opportunity to observe the ice stratification. As the observed layers are frequently separated by dark levels of sediment consisting of dust, clay, leaves, erratic breakdown slabs, tree trunks and branches, relative dating can be carried out. However, Serban et al. (1967) pointed out the difficulty in establishing the time interval corresponding to each layer of ice owing to large melting episodes.

Organic material included in the ice was first investigated by speleologists at the beginning of the nineteen-fifties. Pollen analysis, carried out in the Ghetarul de la Scarisoara, led to an age estimation of 3500 years for the lowest ice layers (Pop & Ciobanu, 1950 [in Silvestru & Boghean, 1992]). This value is consistent with a radiocarbon analysis performed on a wood sample caught in the ice profile (Fanuel, 1993).

Similar investigations were carried out in the Austrian ice caves, where pollen analyses suggested that most cave ice was formed during postglacial or even historic time. Hence, Kral (1968) estimated a maximal age of 500 years for the ice of the Dachstein-Rieseneishöhle which is consistent with estimations based on energy-balance considerations (Saar, 1954 a [in Mais, 1999]). This latter author concluded that the cave ice is not a relict from the last glaciation but was formed around 1300 A.D. Although no radiocarbon analyses were possible in this particular cave, such investigations were achieved in several other Austrian ice caves (Pavuza & Spötl, 1999; Achleitner, 1995). Based on these results, Pavuza (Pavuza & Spötl, 1999; Pavuza & Mais, 1999) concluded that some of the Austrian ice caves (Dachstein-Rieseneishöhle, Eisgruben-Eishöhle, Hundalmeishöhle) were partly or completely free of ice during the post-glacial period (last 10'000 years).

In northern America, radiocarbon dating was done on bones of the Nahanni ice caves and provided a maximum age of 2400 years (Schroeder, 1977). Applying the radiocarbon dating method to cryogenic powder found on the ice, Lauriol & Clark (1993) determined a maximal age of about 6000 years BP in Cavern '85 (Yukon, CA). Analysis of a twig (part of a tree limb) contained in cave ice of Bandera Volcano (New Mexico, US) provided a radiocarbon age of about 3400 years (pers. comm. J. Alford).

Owing to the rarity of adequate material, only few dendrochronological measurements have been attempted. However, successful investigations are being performed currently in Switzerland (e.g. Schlatter et al., 2003) and Romania (Kern et al., 2003).

Determining the age of ice formation remains one of the most relevant questions related to ice caves. If many caves likely enclose only recent ice formations, the investigations carried out on several study sites suggest that millenia-old cave ice may be present. The possibility of Pleistocene age ice fillings constitutes a major motivation for further investigations on ice caves and represents an interesting topic for paleoclimatic research.

2.4.5 Paleoclimatological studies of cave ice

Since working conditions in ice caves are difficult, the selection of an optimal study site is an important factor for glaciological investigations. Because of the subsurface environment, ice corings were only rarely attempted. Furthermore, the seasonal melting observed in most low-altitude ice caves significantly increases the risk of chemical remobilization and complicates the interpretation of analyses. Contamination of superficial ice by recent water (infiltration water, condensation) may also be critical for Tritium analyses. Therefore, the application of this dating method to cave ice could be problematic (Pavuza & Spötl, 1999; Pavuza & Mais, 1999).

Stable isotopes were only occasionally investigated in cave ice. Nevertheless, oxygen and hydrogen isotopes were analysed on different cave ice samples (e.g. Brown & Marshall, 1973; MacDonald & Yonge, 1997; Lauriol & Clark, 1993). Unfortunately, results remain relatively poor at present because of frequent contamination/ remobilization processes. Serban et al. (1967) observed the variations of the isotopic composition (^{2}H and ^{18}O) of various layers in Ghetarul de la Scarisoara and concluded that a periodicity of the minimal values could be observed. Racovita (1972) attempted to correlate the stratigraphic elements from two cores with climate oscillations. In this way, he attributed three impurity layers to well known warm years which correspond to late Holocene temperature oscillations (Serban & Racovita, 1987).

Similar investigations were carried out in the Dachstein-Rieseneishöhle and appear to link layers of increasing thickness with a period of cooler climate. These observations were correlated with a decrease of dissolved carbonates contained in the ice (Pavuza & Mais, 1999).

Unfortunately, at present, the acquired data do not allow a larger synthesis of climatic variations recorded in ice caves. Studying the cave ice archives is a major investigation target for the future. Located in a geographical context where such records are frequently missing (Perroux, 2001), ice caves constitute an interesting alternative for paleoclimatological studies. However, since investigations still have an exploratory character, feasibility studies need to be performed on some selected sites.

2.4.6 Conclusion

Ice caves are mentioned in temperate karst regions all around the world but their distribution seems to be closely related to specific climatic conditions. Although the presence of millenia-old subsurface ice is suspected in several caves, field observations lead to the conclusion that the formation of low-altitude cave ice is still an active process. Therefore, generally, cave ice should not be considered as a paleo-environmental relict. This statement is supported by literature references mentioning the presence of ice caves already during the sixteenth century, suggesting that the presence of cave ice is not related to the maximal cooling observed during the Little Ice Age. Nevertheless, observations made during the last century in different European ice caves point out a general melting of cave ice fillings during the nineteen-nineties.

2.5 Conceptual model of the origin of cave ice based on literature studies

The synthesis of numerous studies dedicated to ice caves during the last centuries highlights two main conditions at the origin of subsurface ice fillings: (1) temperatures below freezing point, i.e. < 0 °C; (2) the presence of H_20 in a solid, liquid or gaseous phase. Although both conditions appear to be essential for the formation and conservation of cave ice; studies of ice caves frequently focused on one or the other aspect but not both of them. These distinct scientific approaches are represented by two different schools, the «climatologists» (e.g. Thury, 1861) and the «geomorphologists» (e.g. Kyrle, 1923). The combination of these two points of view enables one to propose a qualitative description of processes at the origin of ice caves. To demonstrate the particularity of ice caves, a conceptual model of the temperature distribution in karst systems is assumed. It is shown that ice caves are often related to thermal anomalies induced by air circulation. A brief discussion of the origin of water is also necessary.

2.5.1 Cave systems

A cave system is defined as a connection of natural openings large enough for human entry (e.g. Gillieson, 1996). Although this anthropocentric definition has no genetic meaning (Klimchouk, 2004), it presents the advantage of focusing on cavities which provide direct access to field observations.
Caves are formed by different processes in many rock types and unconsolidated sediments. Solution (karst) caves are the most common and occur widely in limestone. Dissolution remains the dominant process, although processes such as erosion and gravitational breakdown may take part in their development. Subsequently, patterns of solution caves are controlled mainly by hydrogeologic factors. Cave systems with great vertical extensions are dominant in mountain regions where thickness of karstified rocks and the vadose zone is great.
Speleogenetic consideration implies that karst conduits can be partly or entirely filled by water, air and sediments during various stages of their development. Cave sediments can either be of allochthonous origin (for instance, Quaternary moraines or deposits related to extreme geomorphic events outside the cave: landslides, storm surges, etc) or autochthonous origin (for instance, breakdown, weathering) (e.g Gillieson, 2004; Sasowsky, 2004). These sediments significantly influence the air drainage of a cave system and thus also the microclimatic characteristics prevalent in the subsurface atmosphere.

2.5.2 The thermal characteristics of cave systems

The high density of conduits present in karst massifs facilitates major water and air circulation. These flows are assumed to be sufficiently high to compensate for the influence of the geothermal heat flux. Based on numerous field observations and measured temperature gradients, Luetscher & Jeannin (2004; Part II, p. 52) suggest that the temperature distribution in caves is controlled by the significant effects of ventilation. This assumption was

Tab. 1a

Cave name	Climate
Cathy (Switzerland)	Temperate
Pleine Lune (Switzerland)	Temperate
Bärenschacht (Switzerland)	Temperate
Longirod (Switzerland)	Temperate
Mutsee (Switzerland)	Temperate
Hölloch (Switzerland)	Temperate
Rasse (France)	Temperate
Spluga della Preta (Italy)	Temperate
Tatras region (Poland)	Temperate
Lechugilla (USA)	Semi-arid
Kievskaya (Uzbekistan)	Semi-arid
KT16 (Uzbekistan)	Semi-arid
Gor Ulugh Beg (Uzbekistan)	Semi-arid
Kijahe Xontjoa (Mexico)	Tropical
Muruk (Papua New Guinea)	Equatorial

Tab. 1b

		Heterothermic zone
Borehole name	**Climate**	**Depth** [m]
FM1 (Switzerland)	Temperate	0–60
FM2 (Switzerland)	Temperate	0–80
Cachot (Switzerland)	Temperate	0–50
Basse Fin (Switzerland)	Temperate	0–50
St-Aubin 2 (Switzerland)	Temperate	n.d.

validated by measurements performed in various deep drillholes in the Jura Mountains (Tab. 1a & b), demonstrating the influence of a cave system on the temperature distribution in an entire massif.

Apart from a superficial zone still influenced by seasonal temperature variations (heterothermic zone), karst systems are characterized by a remarkable temperature stability (homothermic zone). This stability can be attributed to the overall heat capacity of the system, consisting of air, water and rock. However, since the rock volume considered is much more important than the air and water fluxes flowing through the system, its heat capacity prevails on the other elements. Thus, Badino (1995, p. 43-45) concluded that temperature fluctuations of more than a few minutes are controlled by the limestone. Figure 2 shows a four-year monitoring of cave air temperatures at a depth of 700 m below the surface (Bärenschacht, Bern, Switzerland). Temperatures were recorded at 1.5 hour time intervals using a Vemco Minilog (resolution of 0.1 °C). The data illustrate the outstanding temporal stability of cave temperatures. The temperatures are consistent with the exterior MAAT at the same altitude and support the hypothesis that the temperature distribution in vadose karst systems is mainly controlled by air ventilation.

Luetscher & Jeannin (2004 ; Part II, p. 52) propose a conceptual model of the temperature distributions in a karst system (Fig. 3). Below the main cave system, the thermal gradient is controlled by the geothermal heat flux.

Heterothermic zone	Ventilated system		Deep vadose zone		
Depth [m]	**Depth** [m]	**T° air gradient** [°C/100 m]	**Depth** [m]	**T° air gradient** [°C/100 m]	**Reference**
0–70	70–290	0.61	n.d.	n.d.	Luetscher & Jeannin (2004)
0–180	180–230	0.59	–	–	Luetscher & Jeannin (2004)
0–20	–	–	20–600	0.33	Luetscher & Jeannin (2004)
0–100	100–180	n.d.	180–450	0.38	Luetscher & Jeannin (2004)
0–140	–	–	140–755	0.32	Weidmann (pers. comm)
n.d.	> 900	~0.5	–	–	Bögli (1980)
0–100	100–470	0.43	n.d.	n.d.	Valton (pers. comm.)
0–130	130–245	0.52	245–878	0.21	Bertolani (1975)
0–300	> 300	~0.5	–	–	Pulinowa & Pulina (1972)
–	0–300	0.55	–	–	Luetscher & Jeannin (2004)
0–100	100–300	0.78	300–420	0.33	Luetscher & Jeannin (2004)
0–25	25–150	0.1	150–200	0.18	Luetscher & Jeannin (2004)
0–150	n.d.	n.d.	> 150	~0.35	Badino (1992)
–	0–320	0.57	320–900	0.33	Luetscher & Jeannin (2004)
0–75	–	–	75–780	0.25	Audra (2001)

Ventilated system		Phreatic zone		Deep phreatic zone		
Depth [m]	**T° air gradient** [°C/100 m]	**Depth** [m]	**T° water gradient** [°C/100 m]	**Depth** [m]	**T° water gradient** [°C/100 m]	**Reference**
60–440	0.61	n.d.	n.d.	n.d.	n.d.	Luetscher & Jeannin (2004)
80–200	0.55	–	–	200–635	~1.5	Luetscher & Jeannin (2004)
–	–	50–130	0.5–0.8	130–170	1.8	Mathey (1974)
–	–	–	–	50–370	3.5	MFR (2001)
n.d.	n.d.	140–320	0.002	320–360	0.42	Bossy & Zwahlen (2000)

Tab. 1 Temperature measurements in various karst massifs (after Luetscher & Jeannin, 2004). Observations have shown that air, water and rock temperatures are always almost in equilibrium. Most observed gradients vary between 0.4 and 0.6 °C/100 m (*z* is positive with depth; n.d., no data). (a) Measurements carried out in caves show the role of air circulation in the temperature distribution of the vadose zone; (b) data from boreholes show the influence of the geothermal heat flux below the main karst conduit network base level and of air circulation in the vadose zone.

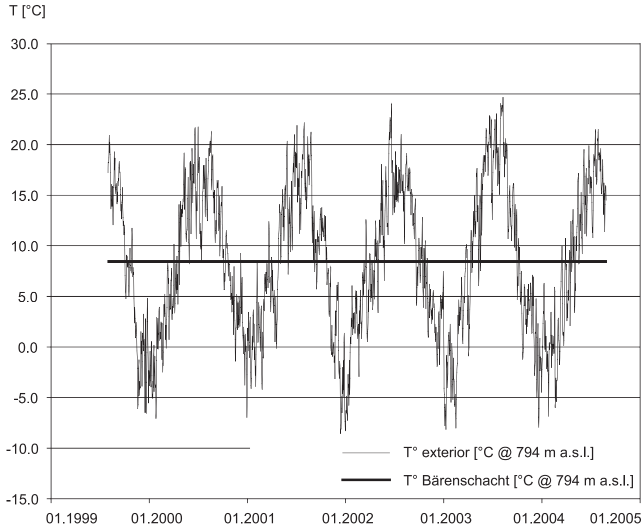


Fig. 2 Air temperature measured at 700 m depth at Bärenschacht (Be, Switzerland). Reconstructed MAAT at 794 m a.s.l. = 8.4 °C (SMA station Interlaken, 580 m a.s.l.); T° Bärenschacht = 8.4 °C (data W. Janz).

Owing to the presence of a major drainage, this heat flux is evacuated by water flow through the main conduit system. Above this base level, the homothermic vadose zone is characterized by a remarkable temporal stability of its temperatures. The temperature distribution in this zone is primarily controlled by the natural gradient of humid air (i.e. 0.5 °C 100 m^{-1}). Near the surface, however, external seasonal climatic oscillations are perceptible (heterothermic zone).

The presence of ice within the homothermic zone is confined to permafrost regions (i.e. T°$_{cave}$ < 0 °C). Due to a deficiency of water circulation (e.g. Ciry, 1962; p. 31) induced by the frozen fracture network, cave ice must often be considered as fossil. Hence, active perennial or seasonal ice accumulations are usually restricted to the heterothermic zone of karst massifs (i.e. active layer in permafrost regions). Favourable to intrusive «ice» (s.l.) deposits (i.e. snow, glacier ice) and subject to seasonal freezing induced by external meteorological fluctuations, «alpine karsts» are likely to enclose such seasonal ice accumulations. However, below the altitude of the 0 °C isotherm, perennial ice conservation is possible only if the cave presents a thermal anomaly. This anomaly is closely related to the air circulation within the system.

2.5.3 Thermal anomalies within the heterothermic zone

Thermal anomalies induced by «chimney effects»

The drainage complexity of a karst massif usually involves significant air circulation through its different conduits. If more than one entrance to a cave system exists, a forced convection can be established resulting from a density difference between the cave air and the external atmosphere (Fig. 4). This chimney effect is well documented by numerous speleologists (e.g. Trombe, 1952; Lismonde, 1981), and similar processes are described in boulder talus (e.g. Delaloye, 2004). Air moisture plays only a secondary role concerning air density in regard to the

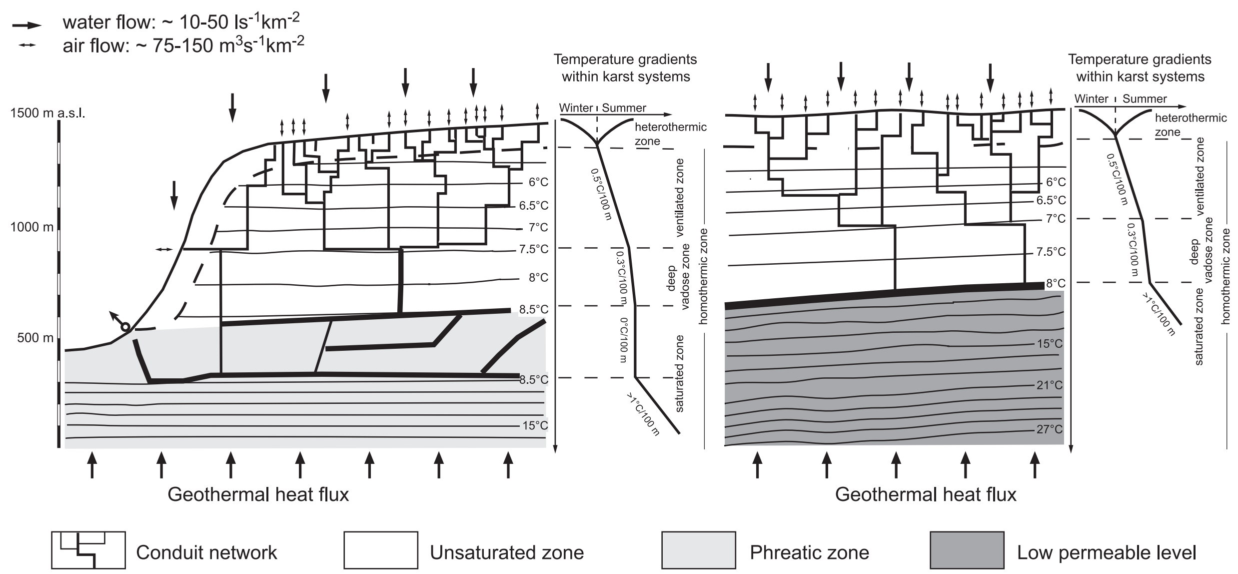


Fig. 3 Conceptual model of the temperature distribution in a karst aquifer (after Luetscher & Jeannin, 2004). Close to the surface (i.e. for the first 50 m) seasonal variations are still observed. This is the heterothermic zone. Highly ventilated conduits located on the top of the unsaturated homothermic zone show steeper temperature gradients than the deep and poorly ventilated part of the vadose zone. Within the saturated zone, down to the bottom of the main conduit network, the gradient is close to zero. Below the main conduit system, temperature gradients are controlled by the geothermal heat flux.

temperature difference. Considering an absence of significant CO_2 concentrations in the cave atmosphere, a physical approximation is provided by Equation (2.1) (Lismonde, 2002):

$$\Delta P_m \approx \rho_0 \frac{p}{p_0} \frac{g}{273} (\theta_{int} - \theta_{ext}) H \qquad (2.1)$$

where ΔP_m: driving pressure [Pa]; ρ_0: mean density of air [kg/m^3]; p: cave air pressure [Pa]; ρ_0: air pressure at surface [Pa]; g: earth acceleration [m/s^2]; θ_{int}: mean cave air temperature [°C]; θ_{ext}: mean surface temperature [°C]; H: altitude difference between both entrances [m].

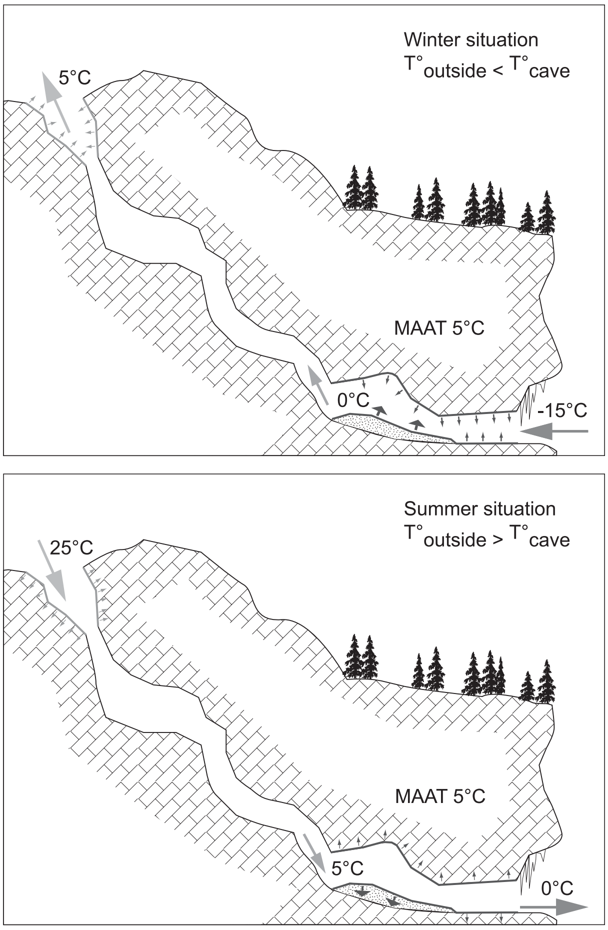


Fig. 4 Schematic behaviour of a chimney effect. During winter season, when the outside temperature is lower than the cave temperature, the light «warm» air of the cave rises up, out of the cave, and is replaced by cold air sucked into the system at the lower entrance. This process is inverted during summer season. As much more energy is stored in the form of ice (latent heat) than in the rock (sensible heat), the temperature anomaly is larger at the lower entrance.

During winter, when the air temperature outside the cave is lower than the cave temperature, cold air is sucked in at the lower entrance of the system ($\Delta P_m > 0$), freezing any existing water infiltrations. During summer, when the exterior air temperature becomes warmer than the cave temperature, the process is reversed ($\Delta P_m < 0$), and warm air is sucked in at the upper entrance. Therefore, a thermal anomaly compared to the MAAT can be observed in both entrance zones, i.e. a cold anomaly at the lower entrance and a warm one at the upper entrance. Hence, below the 0°C isotherm altitude, sporadic permafrost patterns are observed at lower entrances of such cave systems if the required climatic conditions are present.

In a system defined by a segment of a karst conduit and a reference corresponding to the specific enthalpy of dry air at 0°C, it might be possible to observe a perennial ice filling if (2.2):

$$\int_{time} \left(H_{in} - H_{out}\right) dt \leq 0 \qquad (2.2)$$

where H_{in}: specific enthalpy of air sucked into the system during winter; H_{out}: specific enthalpy of air blown out of the system during summer.

A rough estimation, neglecting the effect of evaporation/condensation processes, shows that under the climatic conditions prevailing in the Jura Mountains at least 145 freezing days are necessary for the presence of congelation ice at the lower entrance of a dynamic ice cave. This corresponds in the investigated area to a theoretical altitude of about 1600 m a.s.l.

Owing to favourable speleological prerequisites the distribution of dynamic ice caves is frequent in the Austrian Alps (for instance, Eisriesenwelt or Dachstein Rieseneishöhle). Synthesizing observations made in Dachstein Rieseneishöhle during 30 years, Saar (1954a; 1954b; 1955; 1956; 1957) concluded that ice formation impacted by chimney effects is closely related to exogenic factors (for instance, external temperature) and that endogenic characteristics (e.g. evaporation/condensation, ground heat flux) must be considered only as secondary effects. Therefore, the presence of dynamic ice caves is assumed to be favoured by continental climates characterized by long periods of low winter temperatures.

Thermal anomalies induced by cold-air trapping

In the case of a single entrance, a cavity may act as a thermal trap. Due to density differences, the exchange of air with the outside atmosphere is limited to an «open» period which corresponds more or less to the winter season (when $T°_{surf} < T°_{cave}$) in a descending conduit. In «summer» ($T°_{surf} > T°_{cave}$), natural convection cells are limited to the lower part of the cave and a thermal stratification of the subterranean air is observed, preserving a cool temperature inside the cave ($T°_{cave} <$ MAAT). Under favourable climatic conditions, infiltrated water frozen during winter and/or firn accumulation may subsist during the summer season until the next winter (Fig. 5). Such accumulations of cold air during the winter season remain the most frequent process for the formation of cave ice and might occur at very low altitudes (e.g. Sesiano, 1996). Famous ice caves acting as cold air traps include the Glacière de Chaux-les-Passavents (France) and Scarisoara ice cave (Romania).

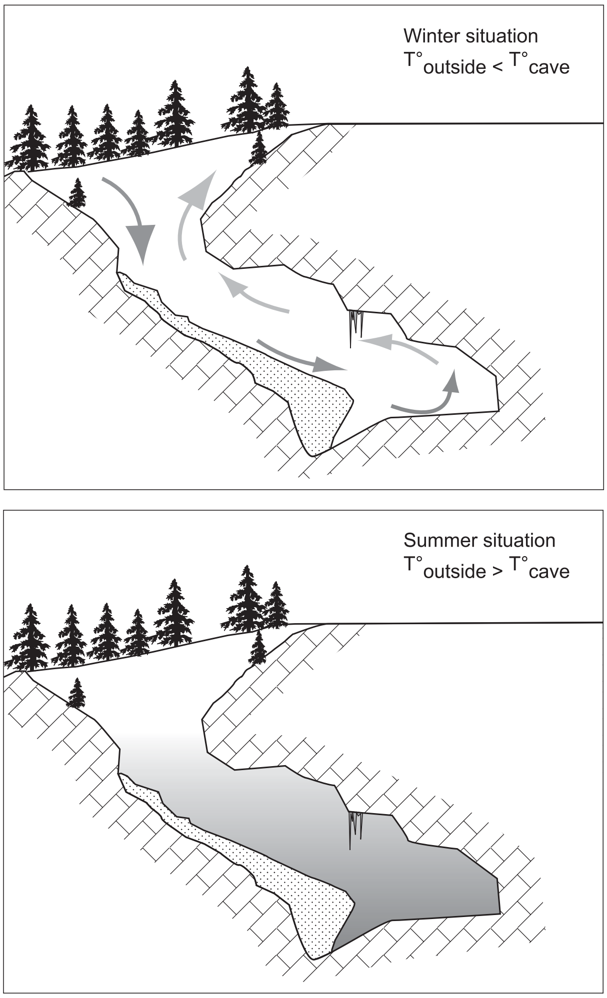


Fig. 5 Schematic behaviour of a cold air trap. During winter season, when the outside temperature is lower than the cave temperature, the light «warm» air of the cave rises up out of the cave and is replaced by the outside colder air. During summer season, due to density differences, cold air is trapped in the cave and a stratification of the subterranean air is observed.

2.5.4 The origin of cave ice

Among many subsurface ice outcrops described throughout the world, at least seven different types of ice are recognized (e.g. Ford & Williams, 1989). Synthesized in Table 2, these types are characterized in two major groups, the exogenous ice (i.e. intrusive snow and glacier ice) versus the endogenous ice (i.e. congelation ice and hoarfrost). Although more detailed morphogenetic subdivisions can be proposed, Luetscher & Jeannin (in press a; Part II, p. 59) point out the difficulty of determining these characteristics in a precise way in the field.

Exogenous ice

Snow (s.l.)

Snow accumulation at cave entrances constitutes the most frequent exogenous ice occurence observed in caves (Fig. 6). Perennial firn accumulations increase with humid climates and are frequent in alpine karst regions. The densification and recrystallization of old snow lead to the promotion of irregular masses with a coarse-grained ice texture. The ice is opaque to blueish in sections. It usually displays distinct layers deposited conforming to the topography and separated from each other by dirt layers (consisting chiefly of external organic material) deposited during catastrophic events in the summer season.

Fig. 6 Subsurface snowdrift (Schwarzmooskogelhöhle/AU). Photo R. Wenger

Intrusive glacier ice

Temperate glaciers covering karst regions can frequently intrude upon karst systems (Fig. 7). Although observations remain rare, such features are sometimes described (e.g. Castelguard cave, Canada; Ford et al., 1976). Frequently admitted in popular interpretations (e.g. Kyrle, 1923; Perrin, 1978), relicts of Pleistocene glaciers have yet to be observed.

Fig. 7 Glacier ice on a karrenfield (Sanetsch/CH). Photo J. Dutruit

Exogenous ice :

Ice type	Description	Formation	Litterature
Firn (recrystallized snow)	Opaque to bluish, layered.	Accumulation of snow in cave traps which densifies and recrystallizes with infiltration component.	e.g. Maire (1990); Bini & Pellegrini (1998).
Intrusive	Massive ice subliming on the caveward face or with hoarfrost formation.	Glacier ice intruded into cave passages at the glacier/rock contact.	e.g. Ford et al. (1976).

Endogenous ice :

Ice type	Description	Formation	Litterature
Drip or flowstone (congelation ice)	Stalactites, stalagmites, flowstone. Polycristalline, clear to opaque.	Freezing of infiltration water.	e.g. Racovita (1994); Pulinowa & Pulina (1972); Kyrle (1923, 1929); Viehmann & Racovita (1968).
Ponded ice (congelation ice)	Clear or coarse polycrystalline ice. Occasionally with bubbles.	Static water that freezes from the top downward. Can incorporate infiltration water and falling hoar ice.	e.g. Marshall & Brown (1974).
Hoar frost	Needles, rosettes and hexagonal plate crystals up to 0.5 m. Small, tapering prismatic crystals.	Humid air condensing onto cave walls below 0 °C.	e.g. Lauriol & Clark (1993); Waldner (1933); Halliday (1966).
Ice in clastic sediments	Intra-particle aggregations as irregular masses, lenses, needles or soil-like extrusions.	Freezing of moist sediments.	e.g. Pulinowa & Pulina (1972); Schroeder (1977).
Extrusive	Curving fibrous crystals, similar to gypsum flowers up to 20 cm long.	Supercooled water forced through microfissures in rock below 0 °C which freezes as it emerges from the cave walls.	e.g. Ford & Williams (1989).

Tab. 2 Perennial ice formations observed in caves (adapted from Ford & Williams 1989; Hill & Forti 1997; Yonge 2003). Subsurface ice patterns can be distinguished between exogenous and endogenous cave ice.

Endogenous ice

Congelation ice (s.l)

In the vicinity of most caves, limestone is fractured and karstified, and therefore subject to significant water circulations. Some of this percolation water might reach the cave, and its freezing can be the origin of major accumulations of congelation ice (Fig. 8). This congelation ice can be found in the form of drip- or flowstones, or as ponded ice. It is clear to opaque and presents a polycrystalline structure with occasional air bubbles.

Fig. 8 Subsurface congelation ice (Eiskogelhöhle/AU). Photo R. Wenger

Hoarfrost

In his physical approach to ice formation processes, Bock (1913) estimated that during the early summer, shortly after the thermal inversion, ice was generated by the freezing of condensation water. The «warm» air, saturated in humidity, flows over the frozen substrata (rock or ice) and is cooled enough to allow condensation (Fig. 9).

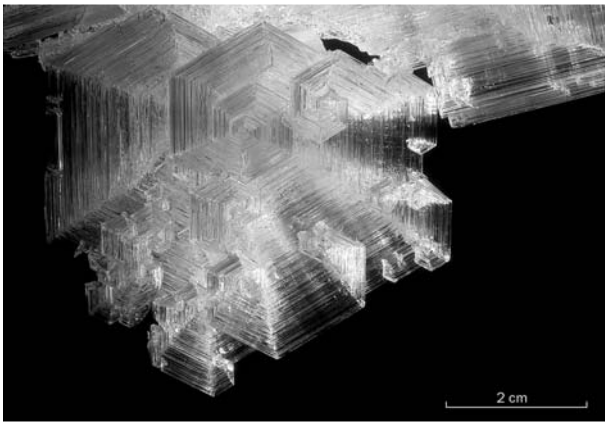


Fig. 9 Subsurface hoarfrost (Pinega/RU). Photo R. Wenger

Seasonal hoarfrost patterns are frequent in the alpine belt, but perennial cave ice resulting predominantly from condensation water is confined to permafrost regions. There, condensation occurs chiefly in cave zones presenting a «warm» thermal anomaly (i.e. upper zone of a multiple-entrance cave system or upwards sloping conduit acting as warm air traps). Condensation ice occurs in the form of needles, rosettes or hexagonal crystals.

Accretion rate is rapid at temperatures close to 0 °C but Ford & Williams (1989, p. 354) estimated that the formation of some large individual crystals requires up to 50-60 years at temperatures ranging between -1 and -5 °C with a very slow, gentle air circulation supplying the water vapour.

2.5.5 A classification of ice caves

The analysis of former studies shows a heterogeneous (if not contradictory) nomenclature related to ice caves (Luetscher & Jeannin, 2001). Nevertheless, the process-based description of ice caves suggests a synthetic classification based on climatological and glaciological observations (Luetscher & Jeannin, in press a; Part II, p. 59). The classification proposed in Figure 10 is based primarily on criteria related to cave-air dynamics and is consistent with the terminology introduced by Thury (1861), which is still in use in most of the English literature. Hence, the terms static, statodynamic and dynamic, refer to the presence or absence of forced convection (static for thermal traps and dynamic for chimney effects). A second criterion involves the ice formation processes (firn accumulation versus congelation ice), somewhat in agreement with ideas reported in French literature (Maire, 1980; 1990).

This classification constitutes a meaningful compromise among former studies and is consistent with the most common terminology. However, it is restricted to caves with massive ice occurrences, i.e. firn and congelation ice fillings, located in low-latitude regions (i.e. «alpine» type). Caves with perennial hoar frost, frozen sediments and ice extrusions are not considered here. Because of the very restricted number of known examples, intrusive glacier ice is also considered as a separate category.

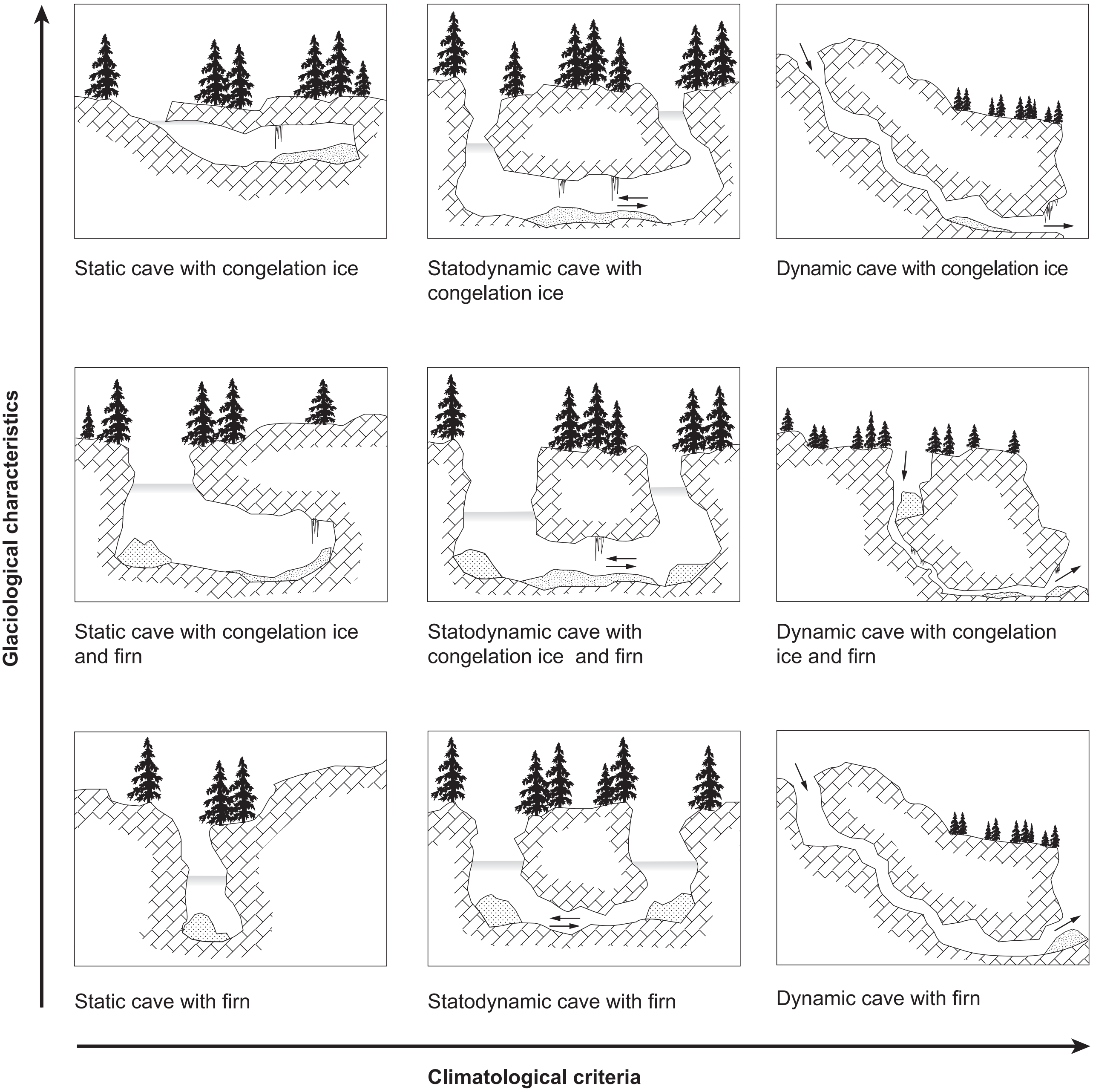


Fig. 10 Suggested classification of ice caves (after Luetscher & Jeannin, in press a). Two criteria are considered in this classification: 1) On the horizontal axis, cave air dynamics distinguishes caves where thermal trapping is at the origin of ice from those where a cold thermal anomaly is induced by chimney effect; 2) On the vertical axis, ice types are distinguished between exogenic ice (i.e. firn accumulation) and endogenic ice (i.e. congelation ice).

3 | Problem definition

3.1 Objectives of the study

Regional literature published in speleological journals well documents individual ice caves, but little data is available to predict the evolution of subsurface ice occurences in a changing climate context. Although an increasing number of studies concern glaciochemical data of cave ice, the preliminary identification of the physical processes at the origin of cave ice is still very limited. The present work intends to improve understanding of the parameters controlling the existence of cave ice and their relationship to external climatic fluctuations. In this way, the study contributes to the identification of:

a Ice caves as an environmental marker;
b Ice caves as an archive for climatic records.

3.2 Questions and strategies

The current study focuses on patterns of the sporadic permafrost occurring in karst environments. In order to identify the effects of fast climate changes, our interest mainly concerns mid-latitude/low-altitude cave ice. At the limits of an equilibrated energy balance, such ice occurences are assumed to react very fast with respect to external climatic fluctuations. Hence, low-altitude ice caves are believed to be good investigation objects for the quantification of energy exchanges between ice caves and their surrounding environment.
Geographically well defined and with the benefit of an extensive speleological documentation, the Jura Mountains are considered to be an appropriate site for a general documentation of mid-latitude/low altitude ice caves. Located in a temperate climatic context and with the highest summits culminating at about 1700 m a.s.l. (MAAT ~3.5 °C), this mountain range does not belong to previously recognized permafrost areas (e.g. Keller et al., 1998).

This study was completed in three successive steps:
1. a detailed inventory was compiled in order to acquire an overview on the distribution of cave ice;
2. a representative process description was developed for a selected study site in the Jura mountains;
3. paleoclimatic potentials of low altitude cave ice were explored.

The following section presents the main issues related to the study of ice caves in the Jura Mountains and the chosen approach:

1. *Ice cave distribution*
Main issues:
– How is cave ice distributed? Is cave ice a frequent feature in the Jura Mountains?
Activity: regional speleological literature was reviewed and a detailed inventory of ice caves in the Jura Mountains was compiled (see 2.1 & 6.1.1).
– Are there some morphological constants which are characteristic of the different ice caves present in the Jura Mountains? What are the observed ice volumes?
Activity: cave descriptions were compiled for most ice caves recognized in the Jura Mountains. Ice volumes were assessed for each visited cave (see 6.1.1).
– What was the evolution of ice during the twentieth century, i.e. in a warming context? Can evidence of former and now vanished ice fillings be found in caves of the Jura Mountains?
Activity: historical documents were collected and cave ice mass balances reconstructed for major ice caves. Cave sediments were observed in order to identify former ice fillings (see 6.4 & Part II, p. 65).

2. *Process identification*
Main issues:
– How does the cave ice accumulate? Is cave ice composed of firn accumulations, congelation ice, hoarfrost, or of all three types?
Activity: the origin of cave ice was determined by field observations; multiple observations also determined the main crystallization periods (see 6.1.3).
– Are subsurface ice accumulations still active or must they be considered as relicts from colder periods? Could cave ice be related to Pleistocene glacial ice? Is cave ice «temperate»?
Activity: ice fluctuations were measured in selected caves and ice temperatures were monitored in a borehole in Monlési ice cave (see 6.1.4 & Part II, p. 77).
– What are the main processes at the origin of cave ice in the Jura Mountains? What is the role of subsurface ventilation on the temperature distribution in karst systems?
Activity: heat exchanges at the boundaries of the system were monitored at one of the selected sites (see 6.2 & Part II, p. 98).
– How is the presence of low-altitude cave ice dependent on external climatic conditions? How will cave ice evolve under different scenarios of climate change?
Activity: the energy balance of a selected study site was quantified and compared with meteorological records (see 6.3 & Part II, p. 65).

3. *Paleoclimatic outlooks*
Main issues:
– What accessible time period can be expected in subsurface ice fillings?
Activity: dating of cave ice was attempted at two selected study sites (see 6.1.4 & Part II, p. 77).
– What kind of information is expected from paleoclimatic records in ice caves?

Activity: glacio-chemical analyses were carried out and interpreted with respect to the identified processes (see Part II, p. 77).

3.3 Definition of the investigated system

According to the proposed nomenclature, «ice caves» are defined as rock-hosted caves containing perennial ice or snow, or both (e.g. Luetscher & Jeannin, in press a; Part II, p. 59). Since the identification of processes at the origin of cave ice relies on a thermodynamic approach, the quantification of heat exchange at the system's boundaries was performed for a representative ice cave.

3.3.1 Global system

The investigated system comprises the entire «thermal anomaly» related to an ice cave (Fig. 11). It includes the ice, the water passing through the cave (snow, water and vapour), the cave air, and the limestone surrounding the cave up to a distance where the rock temperature remains stable throughout the year.

Boundary conditions are defined by a) a constant rock temperature and b) exterior meterological fluctuations above the entrance shafts (daily and seasonal oscillations). Heat exchanges at the boundaries comprise:

1 forced convection;
2 conduction through the surrounding limestone;
3 advection by water circulation;
4 thermal radiation.

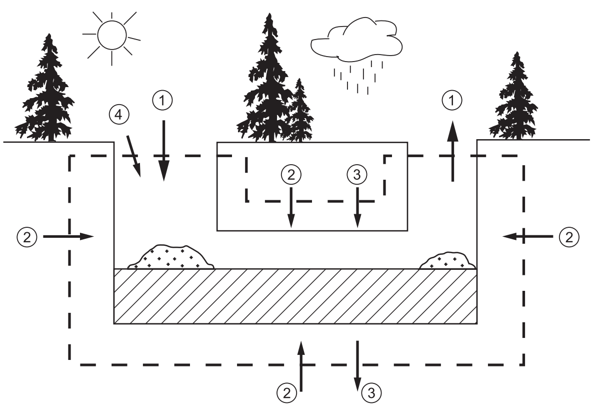


Fig. 11 Definition of a global system describing ice caves. Heat exchanges observed at system boundaries comprise (1) Forced convection; (2) Conduction through the surrounding limestone; (3) advection by water circulations; (4) solar radiation.

The thermodynamic referential is taken as the energy equivalent for melting pure ice. This implies a zero enthalpy for dry air at 0°C, a zero enthalpy for water at 0°C, a positive enthalpy for water vapour and a negative enthalpy for ice and snow (latent heat). The annual energy balance of the global system can be expressed as:

$$\Delta E_{air} + LE_{snow} + \Delta LE_{ice} = S + \Delta E_{water} + R \quad (3.1)$$

Where: ΔE_{air}: heat advected by air circulation; LE_{snow}: latent heat of intrusive snow; ΔLE_{ice}: latent heat of ice; S: ground heat flux; ΔE_{water}: heat advected by water circulation; R= thermal radiation.

ΔLE_{ice} represents the adjusting parameter related to the cave ice mass balance. As many of these heat fluxes are difficult to assess, the global system was subdivided into several smaller sub-systems (air, ice and rock), which could be validated more easily by field measurements. The sub-systems are constituted of more or less constant volumes. Since they can be passed through by air and/or water fluxes they are considered to be open sub-systems.

3.3.2 Sub-system air

The sub-system air is defined by considering only the cave atmosphere of more or less constant volume. Because it can be crossed by air and water flows, it classified as an open sub-system. Heat transfers at boundaries include solar-, rock- and ice- radiation, phase changes issued from evaporation/condensation processes (sublimation and condensation), advected energy related to airflows and advected energy related to water flows. Since radiation affects the rock and ice surfaces rather than the air itself, it is assumed that this term can be neglected in the energy balance of this sub-system. Hence, the annual energy balance is given as:

$$LE_{cond.} + E_{air} + A_{water} = LE_{sub.} + E'_{air} + A'_{water} + \Delta H \quad (3.2)$$

Where $LE_{cond.}$: latent heat related to condensation; E_{air}: heat advected by air into the system; A_{water}: heat advected by water into the system; LE_{sub}: latent heat related to sublimation; E'_{air}: heat advected by air out of the system; A'_{water}: heat advected by water out of the system; ΔH: energy lost by the sub-system in contact with the surrounding rock and ice.

Note that $LE_{cond.}$ and LE_{sub} do not represent all phase changes present in the ice cave, since those are mostly local processes. In fact, these latent heat transfers are distributed on both bodies considered (for instance air/rock, air/ice). Their quantification would require the study of the local thermodynamics.

3.3.3 Sub-system ice

The sub-system ice is formed by a well defined ice (s.l.) body. The system is assumed to have a constant volume and heat transfer occurs only by diffusion. Boundary conditions are given by daily and seasonal temperature fluctuation at the upper and lateral limits, while a constant temperature of 0°C is assumed at the lower limit. Thermal characteristics of ice considered here are taken from handbooks (thermal diffusivity of the ice=$1*10^{-6}$ m^2s^{-1}).

3.3.4 Sub-system rock

The investigated sub-system is composed of massive limestone with a constant volume. The boundary condition is geometrically well defined on one side by the cave wall geometry. On the other side of the massif, the sub-system is delimited by a boundary condition defined as a constant temperature over time. This temperature is taken as the Mean Annual Air Temperature (MAAT) outside the cave, which is about 4.5 °C at the Monlési site investigated in this study. Thermal characteristics of limestone considered here are taken from handbooks (thermal diffusivity of the limestone = $1.13*10^{-6}$ m^2s^{-1}).

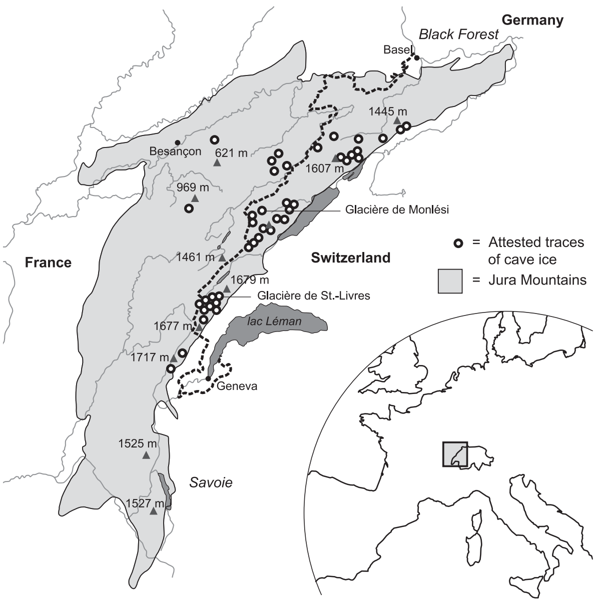


Fig. 12 Geographical situation of the study site. Of over 10,000 caves explored and documented in the Jura Mountains only a few dozens contain a perennial ice filling. These ice caves are mostly located in the inner part of the Jura range, where the highest peaks and the harshest winter conditions prevail.

4 | Selected study sites

4.1 Regional settings (Jura Mountains)

The Jura Mountains form an approximately 40 km wide arc between the Savoie (France) in the southwest and the Black Forest (Germany) in the northeast (Fig. 12). The inner part of this NW-oriented arcuate range, mostly located in Switzerland, is characterized by a succession of crests and valleys ranging between 1000 and 1500 m a.s.l., with the highest peaks reaching about 1700 m a.s.l. The external Jura is characterized by flat plateaus, limited to the North and separated from each other by numerous small-scale tear faults. The mean elevation in this area is about 700 m a.s.l. The Jura Mountains consist essentially of sedimentary rocks (mainly of Jurassic age), presenting a stratigraphy of alternating limestones and marls up to 1000 m in thickness. Dominated by carbonate rocks, most of the Jura Arc is affected by an intense karstification process.

The temperate climate of the Jura Mountains is strongly influenced by oceanic meteorological conditions. As a first obstacle from the Atlantic Ocean, the Jura Mountains benefit from abundant precipitation, usually between 1200 and 1600 mm year^{-1}, which can reach more than 2000 mm year^{-1} on certain crests.

In general, the western side presents wetter conditions than the eastern one, which is already characterized by a semi-continental regime. However, the yearly precipitation distribution over the entire massif seems to be quite random and long-term measurements show a pluviometry varying locally between wetter and dryer regimes (Gaiffe, 2001). Snow precipitation occurs about 50 days year^{-1} on the higher summits and it represents half of the total precipitation observed there. The accumulation can reach more than 2 m and remains from November to May in certain «combes» (small valleys) of the high range.

Owing to karstification, groundwater recharge represents between 50 and 75 % of the total precipitation, depending on the elevation (e.g. Tripet, 1973; Jeannin & Grasso, 1995; Luetscher & Perrin, in press). Most of this infiltration takes place in late autumn and in April during snowmelt. In summer, major precipitation events related to strong thunderstorms can easily represent about 20 or even 50 litres m^{-2} during a short period of time (less than one hour).

Air temperatures present large contrasts between the warmer and the colder months. For instance, the maximal annual amplitude reaches up to 50 °C while a difference of 17 °C between the warmest and the coldest months is common. According to air temperature records (Meteoswiss), an average vertical gradient of -0.5 °C 100 m^{-1} is estimated which is consistent with the wet adiabatic lapse rate (Fig. 13). The altitude of 0 °C-isotherm can therefore be extrapolated to approximately 2200 m a.s.l., which is far above the highest summits.

However, because of its topography, the many closed

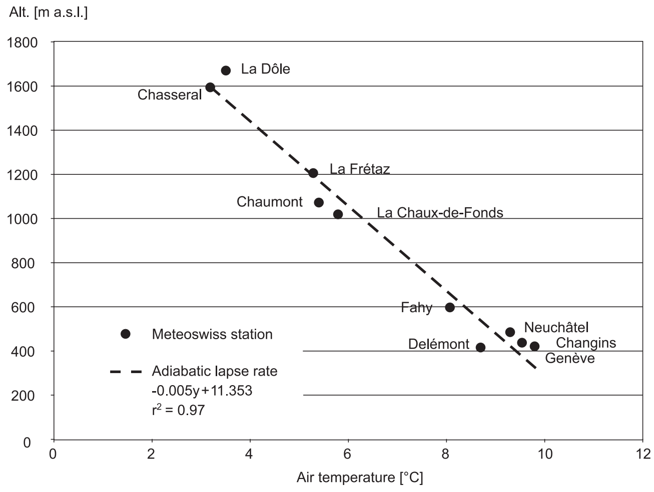


Fig. 13 Temperature gradient in the Jura Mountains computed from Meteoswiss data.

depressions lead to frequent temperature inversions during the winter season. Therefore, it is quite usual to observe warmer temperatures on the slopes than in the valleys.

4.2 The experimental sites

Among the perennial subsurface ice fillings recognized in the Jura Mountains, two ice caves were selected for further investigations: the «Glacière de Monlési» and the «Glacière de St-Livres». Large amounts of massive ice resulting from distinct accumulation processes, visible and accessible ice stratification, and their location in two distinct geographical contexts, constituted the main criteria for the selection of these two ice caves. Furthermore, both sites benefit from extensive documentation by speleologists since the mid nineteen-fifties (e.g. Stettler & Monard, 1960; Stettler, 1971; Gigon, 1976; Dutruit, 1991; Audétat et al. 2002).

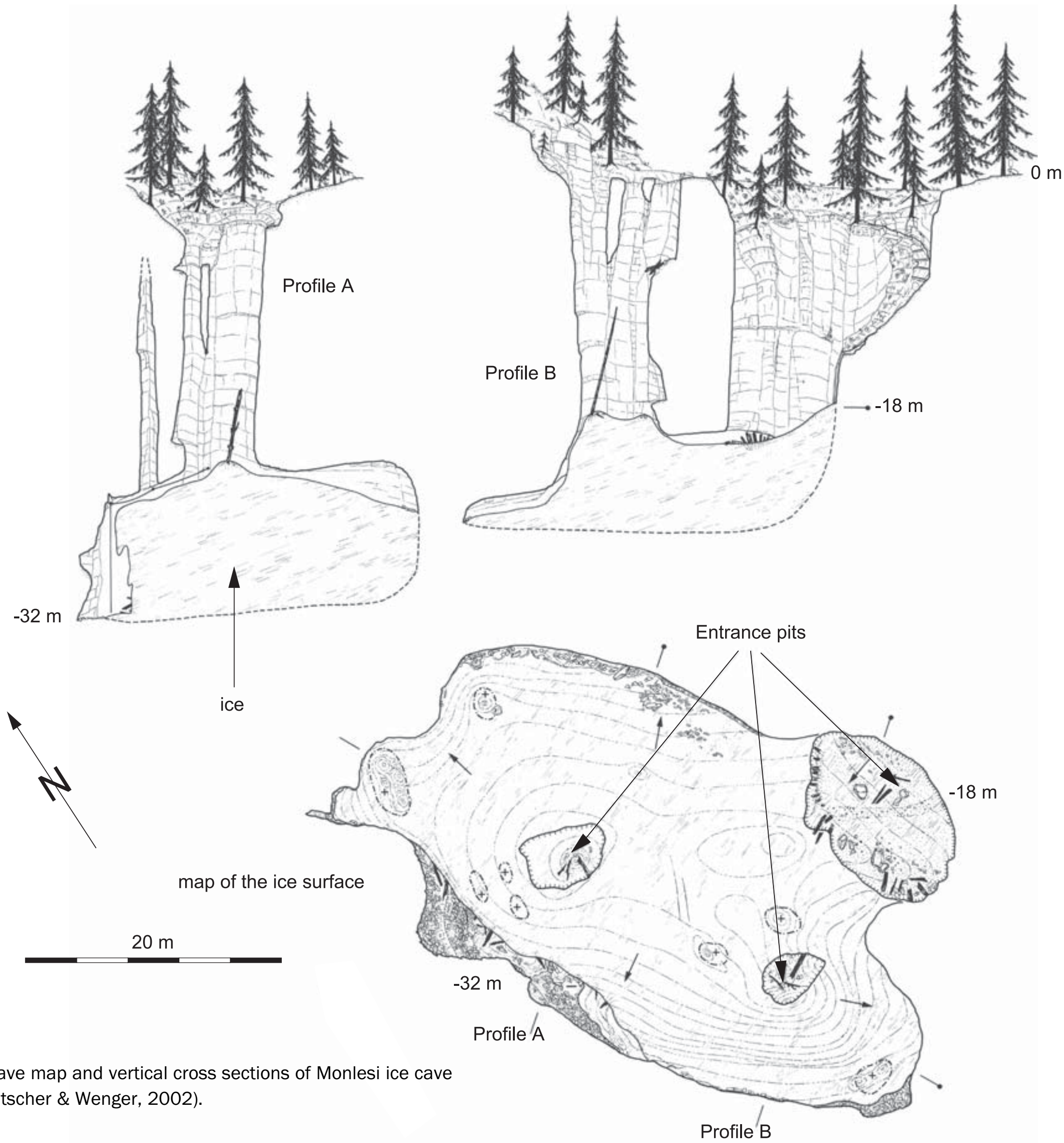


Fig. 14 Cave map and vertical cross sections of Monlesi ice cave (after Luetscher & Wenger, 2002).

Fig. 15 Ice stratification observed in Monlesi ice cave. Photo D. Bolius

Located in the Swiss Jura Mountains (Boveresse, Neuchâtel, 6°35'4''/46°56'18'', 1135 m a.s.l.), the «Glacière de Monlési» is the largest ice cave in the Jura Mountains. Situated in a large closed depression, the cave opens on the edge of a wooded area with three entrance shafts of different diameter, leading at -20 m to a large room of about 20 x 40 x 15 m (Fig. 14). Although all three entrances are almost at the same altitude, significant air circulations can be observed during the winter season. A small oscillating air draft during the summer is considered to be a statodynamic climatic behaviour (e.g. Stettler, 1971) and, therefore, the system is not a true cold air trap. The filling of Monlési ice cave results mainly from the accumulation of annual deposits of congelation ice issued from the freezing of percolation water. The presence of several chimneys close to the surface confines these water inlets to a few infiltration points leading to the formation of major drip- and flowstones. Their formation occurs preferentially during the early spring, when external snow melting brings water infiltrations into a frozen environment. Luetscher & Wenger (2002) assessed the ice volume at 6000 m^3 where morphological evidence suggests a maximal ice thickness of about 12-15 m (Fig. 15). The ice stratification is accessible along a 10 m deep shaft leading to the deepest part of the cave at -32 m. Between the observed ice layers, organic deposits underline a periodicity of the ice formation.

Considered to be the most appropriate site for representative identification of processes at the origin of low altitude cave ice, the Monlési site was selected for the description, the analysis and the quantification of heat transfers observed at the boundaries of the cave.

Contrary to the Monlési ice cave, firn accumulation constitutes a major process at the origin of the St-Livres ice cave. This comparison site is located at 1359 m a.sl. in the western part of the inner range (Bière, Vaud, 6°17'50''/46°33'47''). The cave opens within a grove on the south-facing slope of a major «combe». A large collapsed doline (Ø: ~20 m) forms the unique entrance of this cave and leads to the deepest part of the cavity at -45 m (Fig. 16). The ice filling, estimated at about 1200 m^3, occupies the base of the entrance shaft (Fig. 17). It results essentially from the diagenesis of snow (i.e. firn) accumulated during winter, but local refreezing processes of infiltration water also contribute to the actual ice mass. The development of this subsurface massive ice volume is controlled by a major water inlet located at the extremity of the cave. The discharge rate of this temporary inlet supplies sufficient energy to limit the maximal extension of the ice filling. Major organic deposits can be observed at the ice surface below the entrance shaft. These deposits accumulate preferentially in the summer/autumn seasons during storm events. Here again, the dirt layers delineate the stratification observed at the ice front.

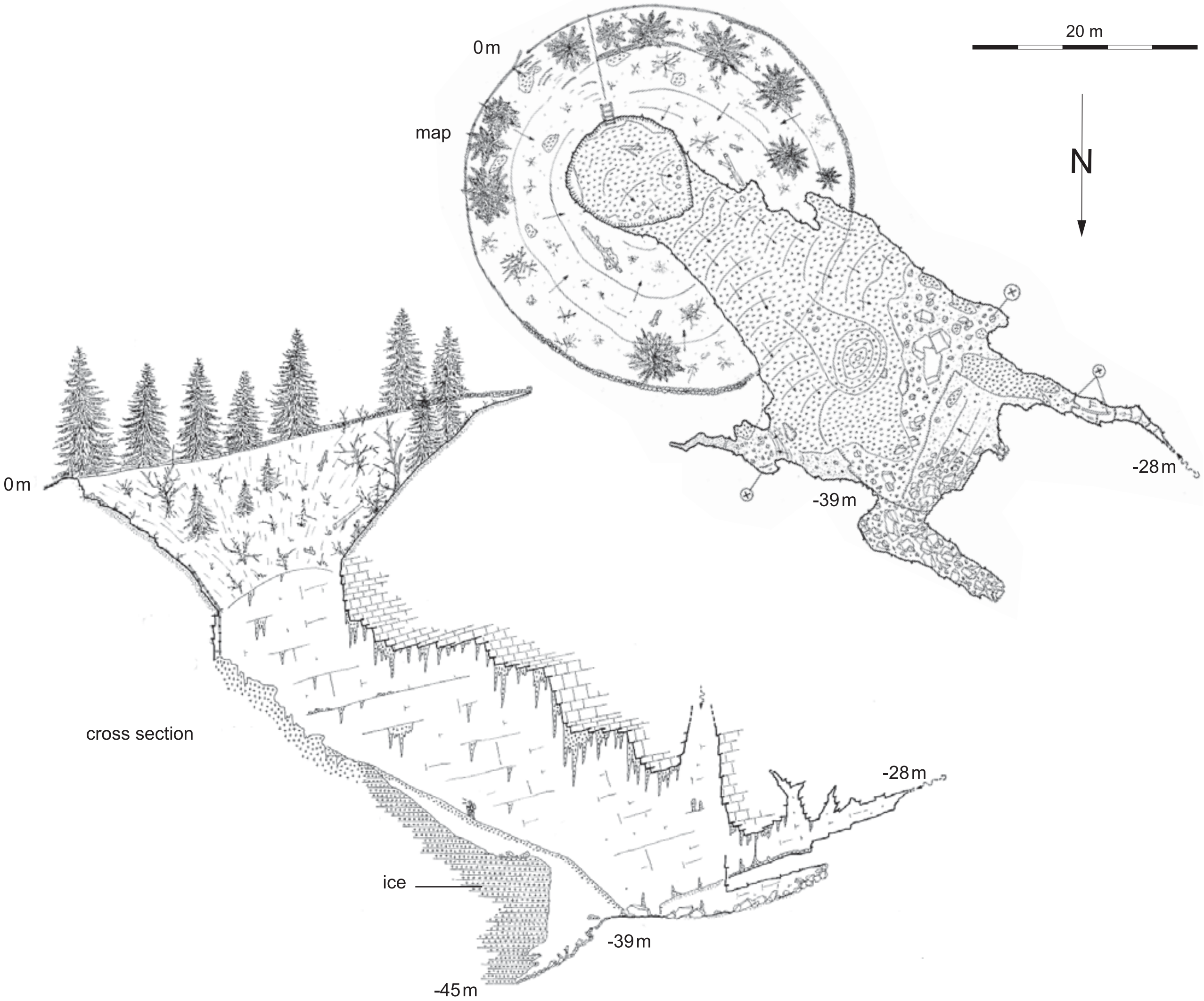


Fig. 16 Cave map and vertical cross section of St-Livres ice cave (adapted from Dutruit, 1991).

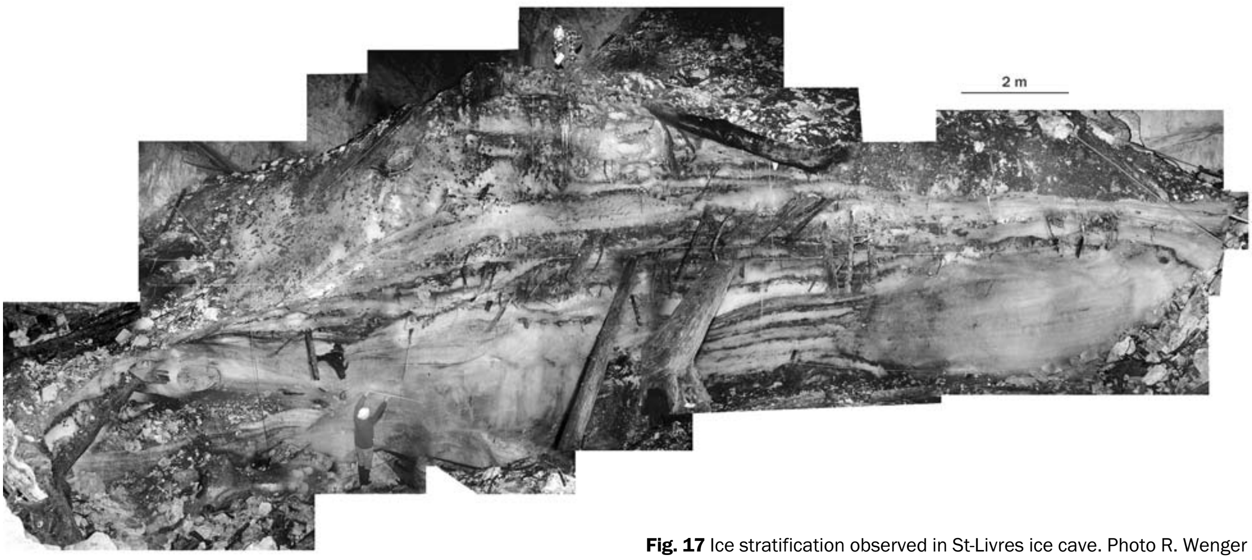


Fig. 17 Ice stratification observed in St-Livres ice cave. Photo R. Wenger

In order to calibrate a physically-based model of heat transfers between an ice cave and the surrounding environment, the quantification of the energy balance was achieved at a selected study site. The measurements required a heavy instrumentation briefly described below.

5.1 Heat transfers – instrumentation, accuracy & validation

The instrumentation of Monlési ice cave included two main stations for the data acquisition, the one being located inside and the other outside the cave. The barely accessible cave environment presented several challenges, including a cave atmosphere saturated in humidity, the presence of freezing temperatures, the abundance of dripping water and various technical problems with the instrumentation, particularly its power supply. The main instrumentation installed on the Monlési site consisted of:

5.1.1 Logging units

The large distances between the measurement stations necessitated the installation of two distinct logging units on the study site. Both units were dedicated to the monitoring of physical parameters involved in the energy balance of the study site:

Exterior

A Campbell CR10X data logger, completed with two multiplexer logging units, was installed in a 745 x 535 x 300 mm polyester box. The power was supplied by a 20 W solar panel (type SX 20 U), a charge controller (Solsum 6.6) and a solar accumulator (12 V, 17Ah). This power unit was dimensioned to provide three-week power autonomy in case of bad weather conditions. The collected data was transmitted by GSM-modem (9.6 kBit) to the office.

Underground

Monitoring performed inside the cave was achieved with a dataTaker DT 500 and completed with a multiplexer 10 c. Power was supplied by a 6 V battery allowing about three month's autonomy at 30-minute time intervals between measurements. The data was recorded on a 2 Mbytes PCMCIA memory card which was changed regularly.

5.1.2 Measured parameters

Precipitations

The Monlési ice cave was equipped in November 2002 with an automated station measuring precipitation outside the cave as well as discharges and temperatures at the main water inlet.

Precipitation was measured using a 400 cm^2 rain gauge station (R32911, Précis Mécanique) connected to the main logging unit. Each tilting corresponds to 0.2 mm precipitation and was summed up once an hour. Owing to technical limitations related to the power supply, it was not possible to use a heated instrument, so the measured data is reliable for rainfalls only (i.e. spring to autumn). Validation is provided by cross-correlation with two nearby MeteoSwiss stations at distances of 5 and 13 km, respectively (La Brévine 6°36'26''/46°58'48'' 1042 m a.s.l. and Les Ponts-de-Martel, 6°43'26''/47°0'18'' 1060 m a.s.l.). Besides local variations in precipitation distribution, the collected data is consistent with general tendencies. Figure 18 shows that measured precipitation is almost 5 percent higher at the study site than at the SMA station (r^2: 0.96; 35 rain events from April to October 2002), a difference which is explained by a higher exposure of the station because of the nearby crest. Although a minor difference in spatial distribution was observed, it is suggested that data recorded at both SMA stations is suitable for the assessment of the pluviometry at the study site:

$$P_{Monlési}=1.05 * [(P_{PDM} + P_{Brev})/2] \tag{5.1}$$

Where $P_{Monlési}$: precipitation measured at Monlési; P_{PDM}: precipitation measured at Ponts-de-Martel; P_{Brev}: precipitation measured at Brévine.

The accuracy is estimated at ± 20 % of the measured values.

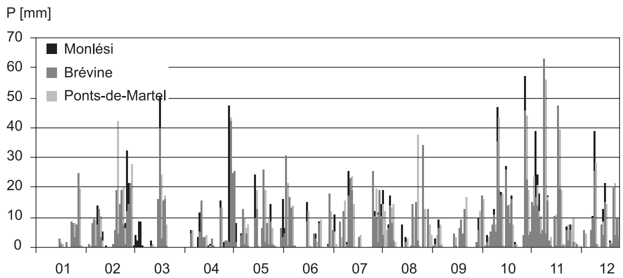


Fig. 18 Precipitations measured on the Monlesi study site versus Meteoswiss data from Les Ponts-de-Martel. Comparison between the different datasets confirmed that local precipitation values could be extrapolated from long term Meteoswiss data.

Water inflow

Temporary water inlets are recognized in numerous places at the Monlési ice cave. All of them are located along tectonic discontinuities leading to open fractures or chimneys. Maximal discharges were assessed during a high-water episode in July 2002 (17.07.02) after 36 hours of continuous rain (56 mm/36 hours). Values vary from less than a few dl min^{-1} to more than 15 l min^{-1}.

Manual gauging achieved during this flood event showed that more than half of the discharge drained through the cave was concentrated in one single water inlet. A station was instrumented to measure discharge rates at 30-minute time intervals. Based on the former experience of

V. Puech (pers. comm.), a pressure probe was set at the bottom of a 1 m long PVC tube perforated by small holes following a helicoidal trajectory. Data was recorded at 10-minute time intervals. The measured water height (H) was converted to discharge values (Q) with a rating curve issued from manual gauging (Fig. 19):

For H<0.5;	Q=14.129*H-3.988	(5.2)
For H>0.5;	Q=30.724*H-12.225	(5.3)

Where Q: [$l min^{-1}$]; H: [m].

Measurements are reliable for discharge values ranging between 1 and 15 $l min^{-1}$. The accuracy is estimated at ±10%. However, because of a frozen water column during the winter season, measurements were mostly confined to the summer season. The advected heat provided by this water was assessed by simultaneously measuring water temperatures with a Pt-100 probe.

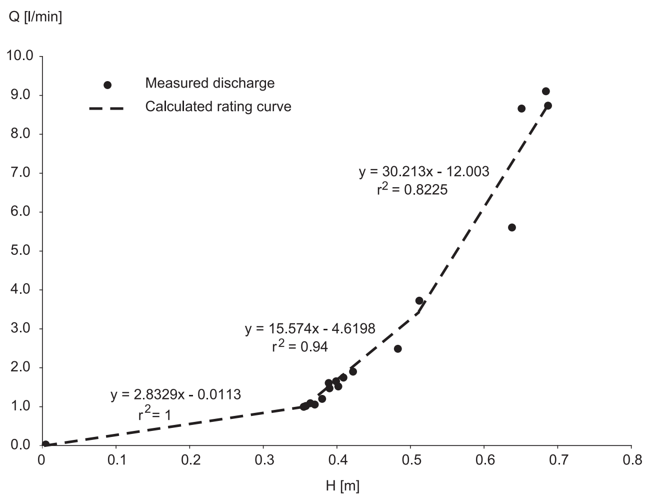


Fig. 19 Rating curve of the main water inlet at Monlési cave. Manual gauging allowed to convert mesured water heights into discharges.

Air temperature

YSI 44006 NTC-thermistors (Negative Temperature Coefficient) were used for measuring cave air temperatures because of their sensitivity to small temperature changes. The measured resistances were converted into temperatures using an empirical mathematical expression for the resistance-temperature relationship. According to the manufacturer's specifications, these thermistors provide an accuracy of ± 0.2 °C at 0 °C, for a resistance of 29490 Ω and a resolution of 1370 Ω/ °C.

The instrumental setting of air temperature measurements comprised two different thermistor chains connected to the logging unit. Consisting of 23 and 21 thermistors respectively, spaced at 2 m intervals, these chains were installed along the two main entrance shafts of the cave and along the cave roof. Air temperatures were recorded at 1-hour time intervals, but short experiments were performed with 30-second time steps.

Additional air temperatures were recorded using UTL-1 miniature temperature loggers (cf. Hoelzle et al., 1999) at 2-hour time intervals. Resolution is given at 0.27 °C.

Rock and ice temperatures

Rock temperatures were measured at 30-minute time intervals in 4 distinct boreholes (80 cm deep) drilled in the cave walls. Pt-100 probes were set at different depths and isolated from each other by polyurethane. In order to improve the heat conduction with the surrounding limestone, each probe was put first into a copper ring having a diameter close to that of the borehole.

Explorative steam drilling (*Heucke ice drill system*) was done to sound the total ice thickness. Drilling performed from the top of the ice filling reached a depth of 8.5 m. Ice temperatures were measured at 30-minute time intervals with five Pt-100 thermistors set at different depths (-1; -2; -4; -5 and -8 m) in the borehole. Owing to percolating meltwater, the borehole refroze within 24 h after drilling.

The accuracy of temperature measurements in ice and rock is estimated at ±0.1 °C.

Accumulation rates & mass turnover

Ice accumulation rates were determined manually by measuring the distance between markers fixed on the rock and the ice surface. An ultrasonic probe (Baumer Electric, UNAM 50I9121/B14) was also installed in order to monitor melting and accretion rates with an elevated time resolution. Accuracy is given by the manufacturer as less than 1 mm.

Mass turnover rates were assessed by measuring the annual displacement of markers fixed within the ice filling. Measurements were performed with the Suunto cave topometry instruments and a digital laser meter. Multiple measurements increase the precision in XYZ positioning to within 1% of the measured distance.

5.2 Glaciological investigation – dating of cave ice

5.2.1 Sampling

Ice coring was performed using the light-weight «Felics» system (Ginot et al., 2002). Owing to the temperate environment, no entire ice core could be extracted. Nevertheless, ice samples of a few cubic centimetres could be extracted down to a depth of 1.7 m (Fig. 20). Some of them consisted of clear ice but others contained numerous organic/sedimentary debris. Thus, ice coring was completed by manual sampling along the ice stratigraphy. This sampling was performed with an electro-mechanical hammer equipped with a hole saw. The five-centimetres thick ice samples (Ø 8 cm) were packed into 0.3-0.8 m polyethylene tubes and transported in dry ice to a cold room kept at -25 °C. Additional water samples were collected manually at the main inlets.

Fig. 20 Ice coring performed in Monlesi ice cave. Owing to the temperate ice features, it was impossible to extract a complete ice core. Neverthelss, ice lenses could be sampled up to 1.7 m depth. Photo M. Luetscher

5.2.2 Analytical methods

Carbon-14 analyses were performed at the AMS-laboratory of the ETH-Zurich by measurement of the $^{14}C/^{12}C$ ratio. Wood samples underwent pre-treatment in a soxhlet apparatus, involving baths of hexane, acetone and ethanol, followed by the standard acid-alkali-acid treatment. The procedure described by Vogel et al. (1984) was applied for graphitization. Calibration was performed using the program CalibETH (Niklaus et al., 1992).

Tritium content (sample volume 10 ml) was determined by direct β^- measurements in a liquid-scintillation spectrometer (Schotterer et al., 1998) at the Physics Institute, University of Bern. The detection limit on a 2σ base (σ=standard deviation) is 1.6 TU.

The ^{210}Pb activity concentration (sample volume 200 ml) was indirectly determined from the activity of its granddaughter nuclide ^{210}Po, electrolytically deposited on Ag plates. The ^{210}Po activity was determined by measuring its α decay at an energy of 5.3 MeV (Gäggeler et al., 1983).

Radon in water samples (20 ml) was determined by liquid scintillation counting (Canberra Packard Tri-Carb 2250CA) at the Centre of Hydrogeology of Neuchâtel (CHYN lab). Cave air samples were collected by passing approximately

Tab. 3 Selected ice caves in the Jura Mountains; 25 perennial ice fillings could be identified in caves (after Luetscher et al., 2005). Perennial ice consists mostly of firn accumulations and sometimes of congelation ice. Estimated ice volumes vary between a few cubic meters to over 6000 m^3 in Monlési ice cave. Evidence of former ice filling could be found in numerous other caves. (n.d.: no available data)

	Name	E	N	Z	Number of entrances	Ice Depth	Cave morphology
1	Glacière de Monlési	6°35'3.5"	46°56'17.6"	1135	3	-20 to -33 m	room
2	Glacière de St-Livres	6°17'46.3"	46°33'55.0"	1362	1	-16 to -45 m	desc. conduit
3	Creux de Glace de Courtelary	7°4'52.7"	47°9'10.5"	1330	1	-25 to -32 m	room
4	Glacière du Crêt des Danses	6°8'6.6"	46°30'12.6"	1490	1	-28 to -34 m	shaft
5	Gouffre de Bellevue	6°40'43.2"	46°54'8.4"	1348	1	-21 to -31 m	shaft
6	Creux-Bastian	6°41'26.8	46°55'4.7"	1210	2	-25 to -27 m	shaft
7	Baume à la Neige	6°18'30.4"	46°35'18.7"	1560	1	-17 to -34 m	room
8	Glacière sud (2) du Mont-Tendre	6°18'59.8"	46°54'8.4"	1575	1	-12 to -15 m	shaft
9	Gouffre du Mont des Verrières	6°28'54.1"	46°53'9.8"	1189	1	-40 m	shaft
10	Gouffre 1 des Grands Bois	6°28'15.8"	46°52'56.0"	1153	1	-20 to -31 m	shaft
11	Glacière de la Pierre-à-Coutiau	6°17'5.7"	46°34'51.9"	1575	1	-16 to -23 m	shaft
12	Creux à la Neige	6°8'47.5"	46°35'37.8"	1410	1	-10 to -15 m	shaft
13	Baume de la Passoire	6°18'37.1"	46°35'22.2"	1550	4	-10 to -30 m	shaft
14	Glacière du Couchant (1)	6°9'48.6"	46°31'4.1"	1470	1	-29 to -31 m	shaft
15	Baume du Bois des Begnines (2)	6°9'40.3"	46°30'55.4"	1510	1	-13 to -15 m	shaft
16	Glacière du Couchant (2)	6°9'37.4"	46°30'21.2"	1430	3	-10 to -14 m	shaft
17	Glacière de Druchaux	6°18'16.8"	46°34'53.8"	1495	1	-30 to -100 m	shaft
18	Glacière de St-George	6°14'25.9"	46°31'33.0"	1290	2	-20 to -25 m	room
19	Gouffre des Croix-Rouge (1)	6°9'34.8"	46°30'51.6"	1520	2	-30 to -35 m	shaft
20	Glacière Paul Matile	6°50'20.6"	47°8'31.2"	980	2	-1 to -3 m	vert. fracture
21	Glacière des Amburnex	6°13'24.5"	46°33'13.8"	1280	2	-4 m	conduit
22	Puits à Neige du Mont Sallaz	6°9'12.5"	46°30'1.2"	1430	1	-5 to -7 m	shaft
23	Gouffre Nord des Cailles	5°57'51.3"	46°17'25.8"	1380	1	-44 to -49 m	shaft
24	Glacière du Col du Crozet (sup)	5°57'51.3"	46°17'25.8"	1545	1	-12 to -25 m	shaft
25	Neigière d'Arc-sous-Cicon	6°24'8.9"	47°1'45.4"	1070	1	-20 m	doline

2 l of air through 180 ml Lucas-cells. The Lucas-cells then were measured at the CHYN-lab (RDA-200, Scintrex). ^{238}U, ^{226}Rn and ^{210}Pb in rock, soil and litter samples (for instance, organic material) were determined by g-spectrometry (HPGe well-type detector, CHYN-lab).

$\delta^{18}O$ analyses of cave ice were carried out at the Paul Scherrer Institute by pyrolisis of the liquid sample at 1450 °C in a glassy carbon reactor to produce carbon monoxide (CO). The $C^{18}O/C^{16}O$ ratio of the gas was measured using an isotope ratio mass spectrometer (Delta Plus XL, Finnigan MAT). Results are reported to the Vienna Standard Mean Ocean Water (VSMOW, Baertschi, 1976):

$$\delta^{18}O\ [‰]=[(R_{sample}-R_{standard})/R_{standard}]*1000 \qquad (5.4)$$

where R is the ratio $^{18}O/^{16}O$ and $R_{standard}$=(2005.20±0.45) * 10^{-6}

Major ions (K^+, Na^+, Mg^{2+}, Ca^{2+}, SO_4^{2-}, Cl^-, NO_3^-) were identified after a 0.45 mm filtration at the Centre of Hydrogeology, UNINE, by ion chromatography (Dionex model DX-120, with a column No. AS14A for anions, and No. CS12A for cations).

6 | Main contributions of the study

6.1 Low-altitude cave ice in the Jura Mountains

6.1.1 Distribution of ice caves

Archeological evidence suggests that the presence of cave ice was known for centuries in the Jura Mountains (cf. 2.1, p. 2), but no synthetic overview of its distribution was available. Historical and speleological data compilation from the Jura Mountains led to the identification of nearly 50 cave descriptions mentioning an ice filling (Luetscher et al., 2005; Part II, p. 65). Although several of these subsurface ice occurrences are limited to seasonal deposits, systematic field validation demonstrated that half of the identified caves still contain perennial ice (s.l.) fillings (Tab. 3).

Mostly located in the inner range of the Jura Mountains, these ice caves are all at altitudes above 1000 m a.s.l. and always consist of downward sloping cave passages generally leading to an obstructed conduit. If more than one entrance is observed, the altitude difference between them is usually within a few meters. Therefore, most ice caves in the Jura Mountains suggest a cold air trap behaviour (e.g. Luetscher & Jeannin, in press a; Part II, p. 59). The only noticeable exception is the Glacière de

Ice characteristics	Estimated ice volume (2003)	Type of ice cave (after Luetscher & Jeannin 2004)	Remarks
congelation ice	~6000	statodynamic with congelation ice and firn	snow at the base of the entrance
firn	~3000	static with firn	stratified ice front of 20 x 7 m
congelation ice	~500	static with congelation ice	
firn	~200	static with firn	
firn	~160	static with firn	
firn & congelation ice	45-60	static with firn and congelation ice	
firn & congelation ice	~50	static with firn and congelation ice	
firn	<50	static with firn	
firn	<50	static with firn	
firn	<50	static with firn	
firn	~40	static with firn	
firn	~20	static with firn	
firn	~20	static with firn	
firn	~15	static with firn	
firn	~10	static with firn	
firn	~10	static with firn	
firn & congelation ice	10	dynamic with firn and congelation ice	
firn & congelation ice	<10	static with firn and congelation ice	
firn	~9	static with firn	
firn	<5	static with firn	
congelation ice	<1	static with congelation ice and firn	
firn	n.d.	static with firn	~15 m^3 in 1989
firn	n.d.	static with firn	~20 m^3 in 1981
firn	n.d.	static with firn	~700 m^3 in 1978
firn & congelation ice	n.d.	static with firn and congelation ice	

Druchaux, situated at 1495 m a.s.l., which shows a chimney effect between the main shaft and a doline located a few meters higher than the cave entrance.

6.1.2 Ice characteristics

Temporary congelation ice (for instance flowstones, stalagmites and stalactites) is frequently found in ice caves of the Jura Mountains. These ice-speleothems are formed by the freezing of a solution issued from percolating drip-water, more or less saturated in calcium carbonate. Through segregation of solutes during freezing, the dissolved carbonate precipitates in the form of a white powder consisting of fine granules of calcium carbonate. In agreement with Zak et al. (2004), it could be observed that this «cryogenic cave calcite» was frequently associated with slow freezing rates. Owing to the feeding drip-water flowing along the speleothem, the small calcite film deposited on the ice surface is often washed away leading to a perfectly transparent ice formation (Fig. 21).

Fig. 21 Transparent ice speleothems in Monlési ice cave. The constant dripwater washed the precipitated cryogenic cave calcite away, leading to a perfectly transparent icicle. Photo R. Wenger

Shumskii (1964; p. 96) observed that if the crystallization rate is fast enough, the calcium carbonate granules remain enclosed in the ice formation as autogenic inclusions, causing a certain opacity to the ice. This stage leads to the characteristic «bamboo» stalagmites which are sometimes observed in caves after a prolonged cold event (for instance in the Monlési ice cave; Fig. 22).

Fig. 22 Bamboo ice stalagmite in Monlesi ice cave. In case of high freezing rates, the calcium carbonate present in the original solution remains trapped in the ice structure, conferring a characteristic opacity to the ice speleothem. Photo R. Wenger

As suggested by Hill & Forti (1997) the warm cave air rising to the ceiling often prevents the formation of icicles. Therefore, ice stalagmites are probably the most common ice-spelothems observed in caves. Kyrle (1923; 1929) demonstrated that the growth of these stalagmites is favoured by fast crystallization rates. This author noticed their increasing diameter when warmer temperatures at the apex of the speleothem cause melting, leading to the so-called «Eiskeulen». When the temperature oscillations imply fluctuating crystallization rates, one can observe on the same speleothem a regular succession of narrow portions of opaque ice and dilated portions of perfectly transparent ice (Fig. 23). Viehmann & Racovita (1968) demonstrated the correlation between this characteristic morphology and the recorded temperature oscillations and emphasized the «thermoindicating» role of such features.

Temporary hoarfrost is observed in numerous cave entrances where warm air flows over a frozen substratum. The freezing of condensation water leads to single hexagonal ice crystals of millimetric to centimetric dimensions. Observations in

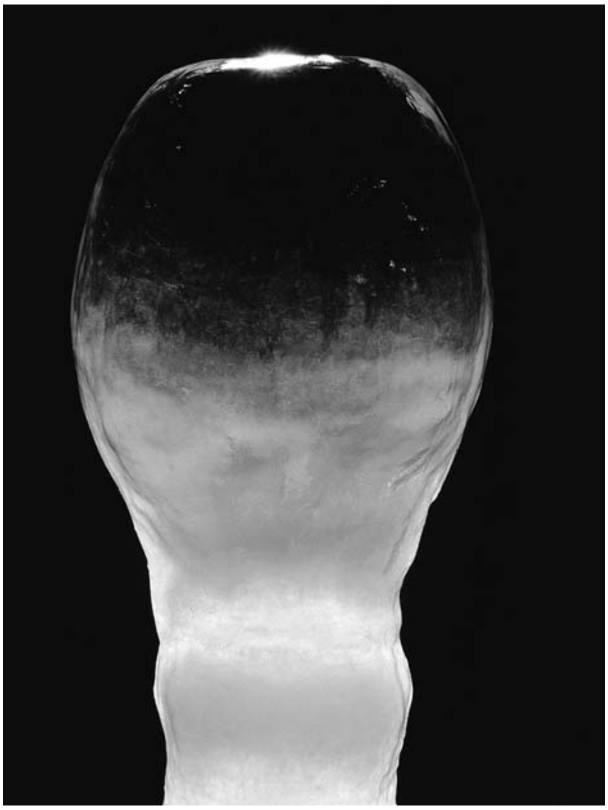


Fig. 23 Thermoindicating stalagmites in Monlesi ice caves. As demonstrated by Viehmann & Racovita (1968), narrow and opaque portions can be attributed to high freezing rates, while large and transparent portions are related to slower freezing rates. Photo R. Wenger.

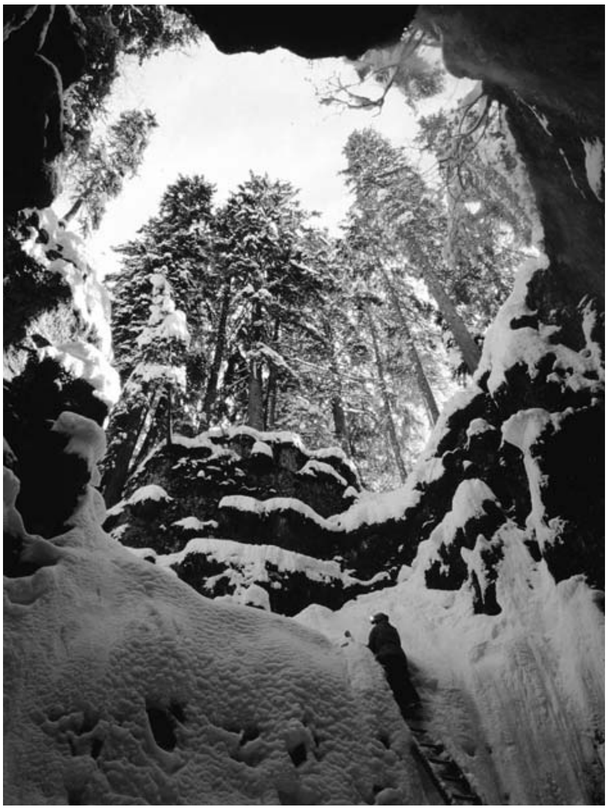


Fig. 24 Snow accumulation at the base of the entrance shaft of Monlési ice cave. Photo R. Wenger

the Monlési ice cave demonstrated that this hoarfrost is at the origin of significant frost weathering of the cave roof, leading Pancza (1992) to attribute a major role of this process to the speleogenesis of the cave.

Field observations demonstrated that perennial ice in caves of the Jura Mountains results from the diagenesis of snow accumulated during the winter season and, more rarely, from massive congelation ice. Even though chemical differentiation of the two is possible (Luetscher et al., in preparation; Part II, p. 77), the distinction between both lithologies was based mostly on genetic processes which are often determined by simple field observations. Most perennial ice fillings observed in caves of the Jura Mountains consist of snow accumulations (Fig. 24), but a noticeable exception concerns the Glacière de Monlési where the ice results essentially from congelation ice (Fig. 25). However, complex sedimentary profiles showing an interstratification between firn and congelation ice are frequent. Observations at the St-Livres ice cave showed that congelation ice consists of thin-layered centimetric ice crystals attributed to individual freezing events. Owing to their strong light absorption, such congelation ice layers are clearly distinguishable. Conversely, firn ice has an anisotropic structure consisting of coarse grained elements. Firn layers are mostly parallel to the substratum and contain frequent organic clasts which are interpreted as seasonal (summer) deposits. These deposits are generally attributed to annual strata but observations also indicate the presence of major sedimentation hiatuses (for instance at Glacière de St-Livres).

Fig. 25 Congelation ice in Monlesi ice cave. The maximal freezing of dripwater is reached when daily melting of the external snow cover is followed by cold nights. This phenomenon is observed particularly during the early spring. Photo M. Anders

6.1.3 Ice dynamics

Freezing of cave ice in the Jura Mountains is favoured by mid-winter warm spells or by the thawing of the external snow cover in early spring. Owing to the reduced actual infiltration observed during these periods, heat exchange with the cave atmosphere is highly efficient, leading to the crystallization of significant amounts of ice. Estimation of seasonal ice accumulation in the Monlési ice cave provided values ranging between 100 and 500 kgm^{-2}, i.e. a thickness of ~10 to 50 cm of ice per year (Luetscher et al., 2003; Part II, p. 122).

However, because freezing/thawing processes slightly modify the shape of the drainage at the water inlets, ice geometries can vary over time. This situation was particularly well set in evidence by comparing the two topographic surveys performed in the Monlési ice cave in 1959 (Stettler & Monard, 1960) and 2002 (Luetscher & Wenger, 2002). Although the overall ice volume decreased between the surveys, a formerly small room is now completely filled with ice and is no longer accessible today (Fig. 26).

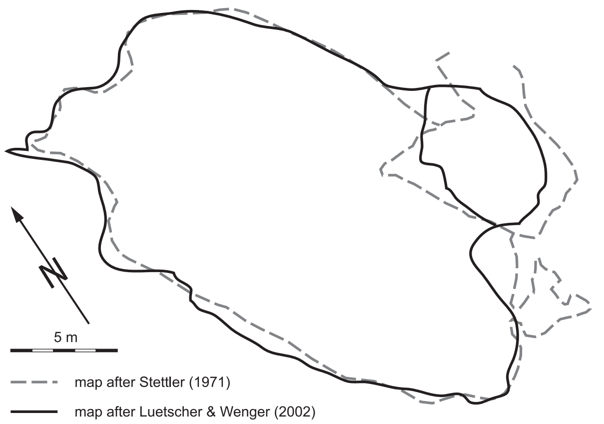


Fig. 26 Comparison between two topometric surveys performed in Monlesi ice cave in 1959 and 2001 respectively. Although a significant ice decrease could be observed, a small room accessible in 1959 is no more visible today. This observation suggests major fluctuations in the cave ice geometry.

Considering that freezing temperatures are still observed during spring, the crystallization continues until a thermal equilibrium is reached within the cave. Field observations helped to estimate that this stage occurs about mid-June in Monlési ice cave. From then, every bit of heat entering the cave contributes to the melting of the cave ice. Since air exchange with the external atmosphere is strongly reduced during the summer season due to cold air trapping, seasonal melting is generally controlled by the percolating water (e.g. during thunderstorms). Although the energy supplied by high discharge rates induces the progressive melting of smaller ice formations, impacts on the global mass balance remain almost insignificant. But owing to the significant heat transfers observed throughout the year at the ice-rock interface, subsurface ice accumulations are subject to noteworthy flowing (e.g. Fig. 27).

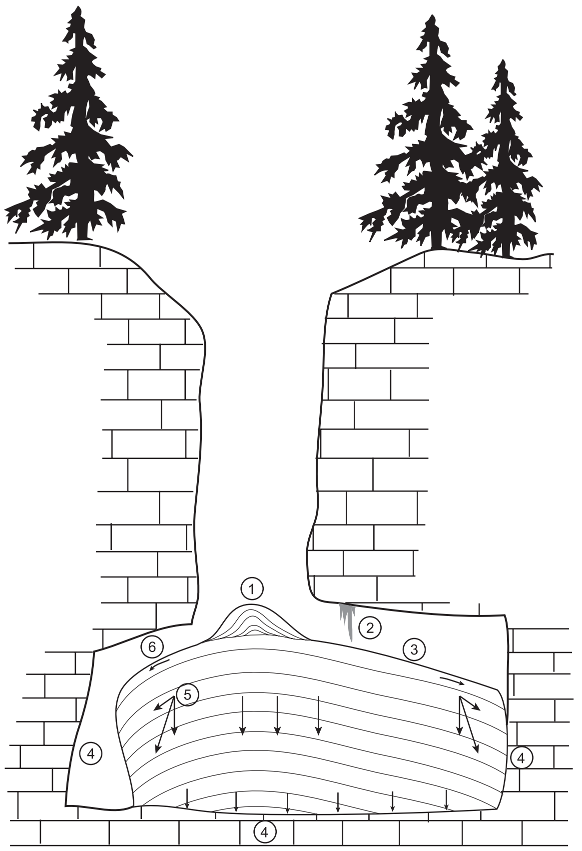


Fig. 27 Accumulation scheme of cave ice in Monlesi ice cave (after Luetscher 2004). 1. Winter snowdrift at the base of the entrance shafts; 2. Infiltration of drip water during the exterior snowmelt; 3. Crystallisation of a new ice layer; 4. Melting of cave ice at the ice-rock interface; 5. Flowing of the cave ice; 6. Superficial melting of the cave ice during the summer season.

Fig. 28 «Neotectonic» signatures observed on ice columns. Horizontal fractures show a displacement in direction of the cave wall, underlining the ice flow induced by melting at the ice-rock interface. Photo R. Wenger

This consideration is well illustrated by the presence of tension cracks observed in ice columns (Fig. 28). Furthermore, the heat flux issued from the underlying karst system induces basal melting, which leads to a constant mass turnover. However, because of the heterogeneity of karst systems, this heat flux changes from one location to the next and the turnover duration can vary significantly between various ice caves. In order to quantify this heat exchange, dating of cave ice was attempted for individual study sites.

6.1.4 Age of the cave ice

Dating of cave ice at the St-Livres and Monlési study sites was achieved by using a multi-parametric approach involving clast-typology, dendrochronology, ice-flow measurements and radiogenic nuclides (Luetscher et al., in prep.; Part II, p. 77). Although industrial nails taken from the upper layers of the St-Livres ice cave confirmed the presence of relatively young ice, dendrochronological datings performed on clastic tree trunks suggested that major hiatuses could be observed in the cave ice stratigraphy (Schlatter et al., submitted). The presence of ice layers older than 1200 years was confirmed by radiocarbon analyses completed on four samples from different stratigraphic layers (Tab. 4). Unfortunately, sampling resolution did not enable a better dating of the identified time gaps, but a detailed stratigraphy provides helpful information on the sedimentation sequence (Kazmer & Luetscher, in prep.).

Figure (29) synthesizes the data from the Monlési ice cave. Results suggest a maximal age of about 120 years. Although the analysis of a twig from the lower ice layer provides an AMS-^{14}C age of 230 ± 45 years BP, a calibrated historical age of the cave ice cannot be proposed due to the temporal variation of the ^{14}C production during the last 300 years. Nevertheless, it is very likely that the sample is less than 500 years old. This upper limit was confirmed by the dating of a tile manufactured between 1874–1916 (pers. comm. B. Boschung, SPMS-Neuchâtel) which suggests a maximal age of the Monlési cave ice at about 130 years. Stable isotope analyses perfomed on samples from the ice core yielded results which were interpreted as being related to fractionation processes induced by air circulation; $\delta^{18}O$ fluctuations of cave ice are probably a good seasonal marker (Fig. 30). Although multi-annual melting events could significantly disturb the counting of individual layers, this method is assumed to be a valuable complement to the dating of a cave ice core with an annual resolution. Valuable qualitative dating was provided by radiogenic dating methods. Analyses performed on fifteen ice samples confirmed the presence of tritium in the upper layers of the ice volume, suggesting an age of less than 50 years. Conversely, samples from a depth of over 5 m do not show any significant tritium content, suggesting an age of over 50 years for the lower part of the ice body. An upper limit is given by the presence of ^{210}Pb in the lower ice samples suggesting an age of about 80 years. Although ice analyses emphasized the risk of contamination related to regional soil enrichment in radium, the study demonstrated the reliability of radiogenic dating methods if all parameters are controlled (Luetscher et al., in preparation; Part II, p. 77).

Tab. 4 Carbon-14 analyses performed on wood samples from Monlesi and St-Livres ice caves. A major time gap is set in evidence in St-Livres ice cave.

Cave	Stratigraphical depth	AMS-^{14}C Age [years BP]	^{13}C [‰]	Calibrated age [BC/AD]		
Monlési	-12 m	230 ± 45	-26.9 ± 1.2	AD	1518 – 1595	(11.3%)
				AD	1620 – 1694	(37.6%)
				AD	1726 – 1813	(40.8%)
				AD	1918 – 1949	(9.2%)
St-Livres	-2 m	190 ± 45	-25.1 ± 1.2	AD	1643 – 1707	(23.8%)
				AD	1719 – 1821	(50.5%)
				AD	1827 – 1884	(10.2%)
				AD	1913 – 1950	(15.5%)
St-Livres	-3 m	1040 ± 50	-25.5 ± 1.2	AD	890 – 1056	(90.6%)
				AD	1088 – 1122	(5.8%)
St-Livres	-4 m	970 ± 45	-24.6 ± 1.2	AD	990 – 1163	(98.1%)
				AD	1172 – 1184	(1.9%)
St-Livres	-4.5 m	1200 ± 50	-25.6 ± 1.2	AD	706 – 754	(12.2%)
				AD	757 – 903	(74.1%)
				AD	916 – 963	(11.0%)

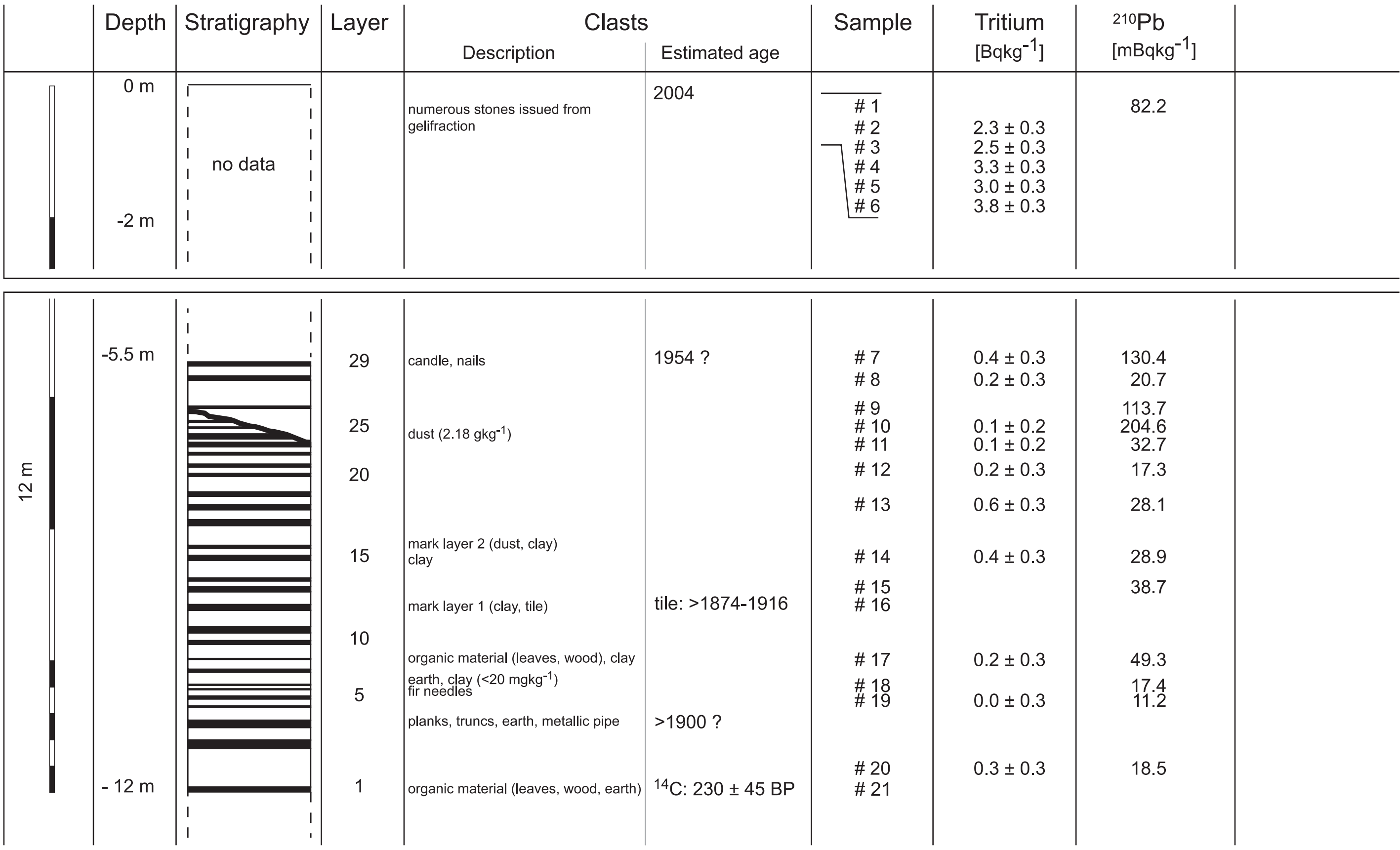


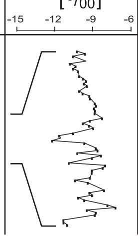
Fig. 29 Synthesis of glaciological investigations achieved in Monlesi ice cave. A multi-parametric dating-approach suggests a maximal age of ~120 years for the lowest ice layers.

Validation of the dating was provided by a multi-annual topometric survey of the ice flow. Results show a basal melting rate of nearly 10 cm year^{-1} (Tab. 5). Considering an accessible ice thickness of about 12 m, this value corroborates the maximal age of 120 years suggested by other methods. Thus, a mean annual ground heat flux of ~1 Wm^{-2} could be assessed at Monlési ice cave.
Dating of cave ice from the Monlési and St-Livres study sites confirmed the presence of two very different mass-turnover rates. This difference is mainly attributed to local ventilation features leading to reduced heat exchanges with the surrounding karst system at St-Livres ice cave.

6.2 Process identification – a case study from the Monlési ice cave

6.2.1 Cooling of the system by air circulation

Field observations performed in the Monlési ice cave document the presence of a significant unidirectional air circulation on cold winter days (Fig. 31). Cold air masses are sucked in through the larger entrance shaft, warm up in contact with the ice and the cave walls, and are blown out through the two smaller shafts. Conversely, during the summer season this ventilation is reduced to a small oscillating airflow. These observations demonstrate the presence of two distinct phases corresponding to an open and a closed system, suggesting a thermal trap behaviour. Figure (32) illustrates a three-year temperature record from the ice cave. This data highlights a period ranging from November to April, which is characterized by a good correlation (r^2=0.91) between the cave air temperature and the negative external air temperature. This signature is interpreted as the result of significant heat exchange related to air circulation. Luetscher & Jeannin (in press b; Part II, p. 93) estimate that the system was open for about 1150 hours during the 2002-2003 annual cycle.
Beyond the near proximity of direction inversions, air-flow measurements performed in the Monlési ice cave provided values varying between 1 and 15 m^{3}s^{-1}. In the largest parts of the cave (D ≈ 5 m), air velocities were assessed to vary between 5 and 80 cms^{-1}. Subsequently, the airflow in the ice cave is considered to be fully turbulent and in the fully rough zone if the airflow is higher than 2.5 m^{3}s^{-1}. Since the roughness of the cave appears to be mostly

$\delta^{18}O$ [$^0/_{00}$]	Age [year AD]	Interpretation
	2003 ~1985	**Stable isotopes**: the oscillating $\delta^{18}O$ record is attributed to fractionation processes induced by winter air circulations. Hence, stable isotopes in cave ice can be considered as good proxies for winter temperature fluctuations.
	~1950 ?	**Tritium**: the very low tritium content of ice samples issued from the lower part of the stratigraphy suggests an age of over 50 years.
		Lead-210: the presence of ^{210}Pb in the lower layers suggests an age of less than 250 years. Owing to the contamination induced by radon enriched water circulations, input activity is not controlled. However, an age ranging between 80-120 years can be assessed.
	~1885 ?	**Clasts**: a calibtrated ^{14}C analyse suggests a maximal age of 500 years. The presence of anthropogenic material confirms that cave ice could even be less than about 130 years old.

Tab. 5 Basal melting rate observed on a three year topometric survey in Monlesi ice cave. Measurements suggest a ground heat flux of ~1Wm^{-2}, leading to a cave ice mass turnover rate estimated at about 120 years^{-1}.

	Distance	Azimuth [G]	Angle [G]	Height [m]	ΔH [m]
08.10.2001					
nail 1	3.41	61	22	1.16	0.00
nail 2	3.28	61	11	0.56	0.00
nail 3	3.26	61	0	0.00	0.00
20.11.2003					
nail 1	n.d.	61	18		
nail 2	n.d.	61	7		
nail 3	n.d.	61	-2.5		
23.03.2004					
nail 1	3.34	61	18	0.93	-0.22
nail 2	3.42	61	7	0.38	-0.19
nail 3	3.37	61	-3	-0.16	-0.16

controlled by the lithology, a mean value of 50 cm appears to be reasonable, leading to a headloss coefficient of ~0.2. Airflow measurements achieved in the ice cave (Fig. 33) confirmed the linear correlation (r^2: 0.99; 9 values) between measured air velocities and the square root of the temperature difference between the exterior air and the cave air. The computed aeraulic resistance of 0.04 kg^{-1}m^{-1} is considered reliable.

Hence, the airflow circulating through the Monlési ice cave was reconstructed from the temperature measurements. The computed air volume drained through the cave during the 2002-2003 annual cycle reached nearly 14*10^6 m^3. Luetscher & Jeannin (in press b; Part II, p. 93) assessed the equivalent airflow during the open period at about 3 m^{3}s^{-1}. Disregarding phase changes, the sensible heat exchange within the cave is estimated at -43 GJ.

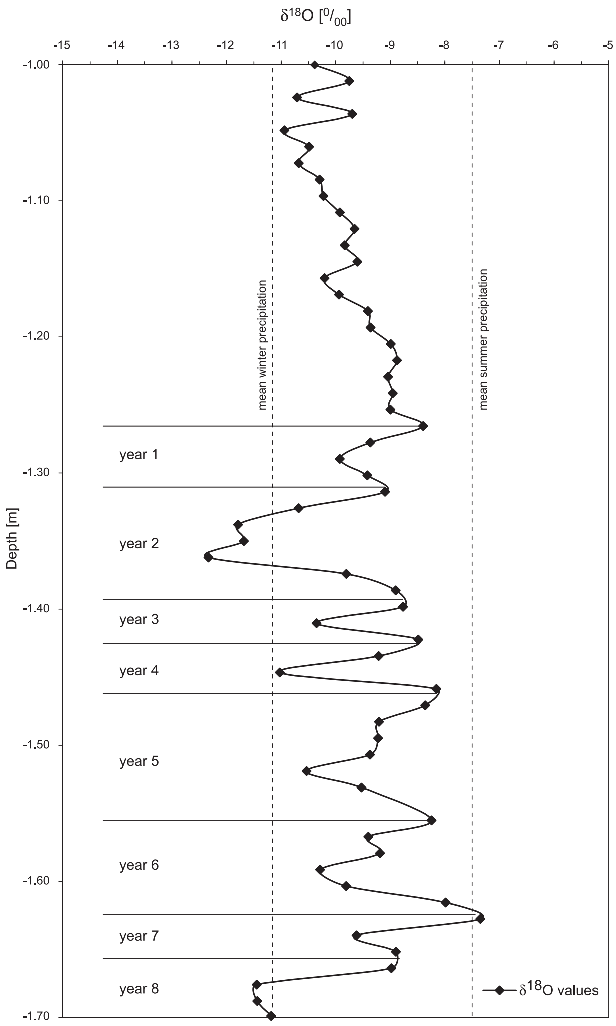


Fig. 30 $\delta^{18}O$ values in ice samples issued from Monlesi ice cave at a depth ranging between -1m and -1.7m. Oscillations are attributed to fractionation processes related to winter air circulation.

Based on the assumption that the external relative air humidity at the site is more or less equal to that measured at the nearby MeteoSwiss station of La Chaux-de-Fonds, and that outflowing air is saturated in humidity, the total heat exchanged by forced convection is estimated at -82 GJ during the 2002-2003 annual cycle. In the same way, the thermal effect of evaporation was calculated to be 39 GJ, the same order of magnitude as the energy corresponding to pure sensible heat transfer. Figure (34) shows the humidity of the outflowing air measured during a three-day survey in winter 2004. The data underlines values close to saturation during the closed period (i.e. Q_{air}: ~0 m^3s^{-1}) but demonstrates that air circulation significantly reduces the specific humidity of the cave air. Hence, the computed effect of evaporation on the energy balance is considered to represent an upper-bound value.

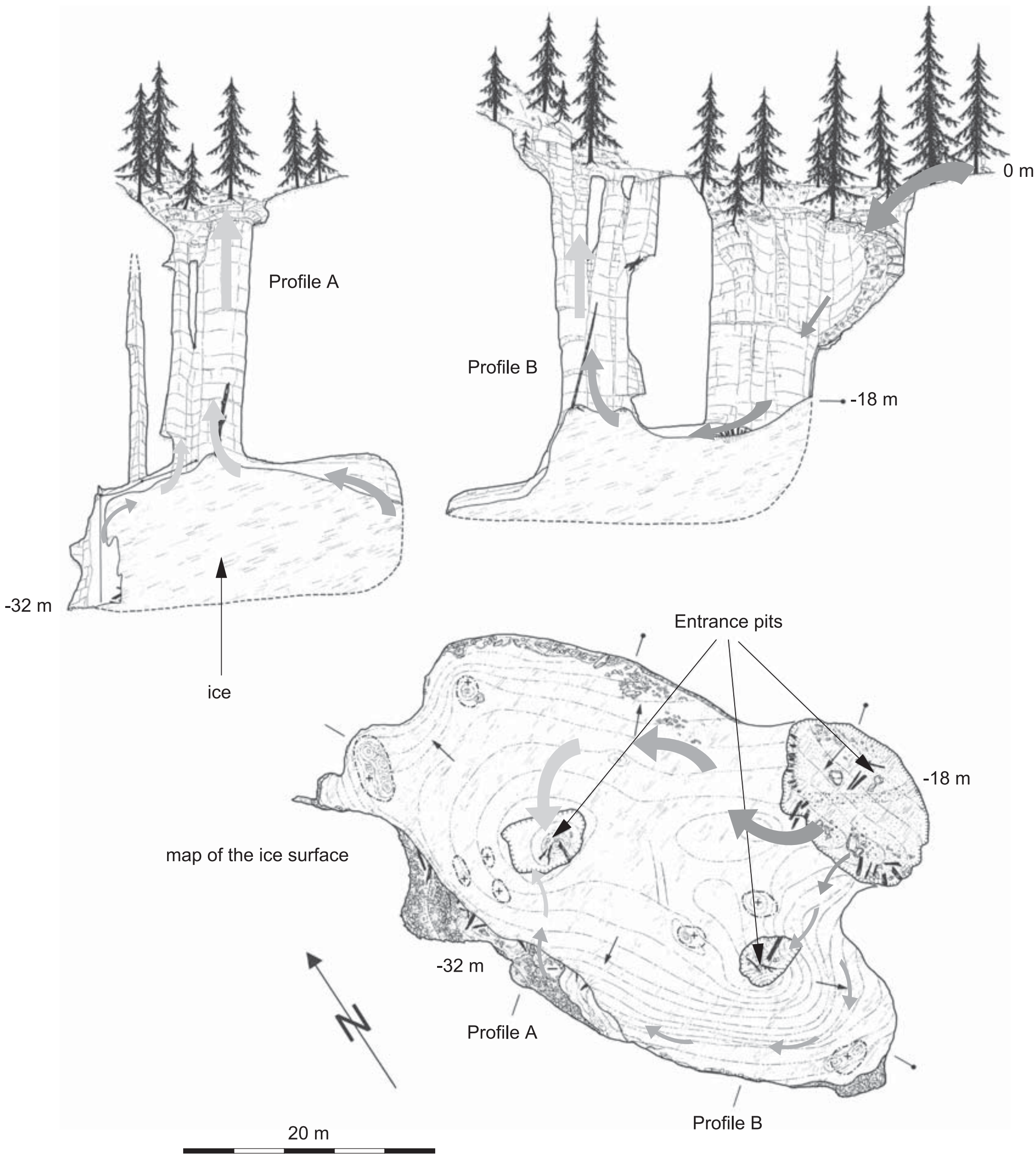


Fig. 31 Plan view and vertical cross sections of the «Glacière de Monlési» (adapted from Luetscher & Wenger 2002). Air circulations during open period are schematised by the arrows. Cold air drops inside the main shaft, warms up in the cave and is blown out from the two other shafts.

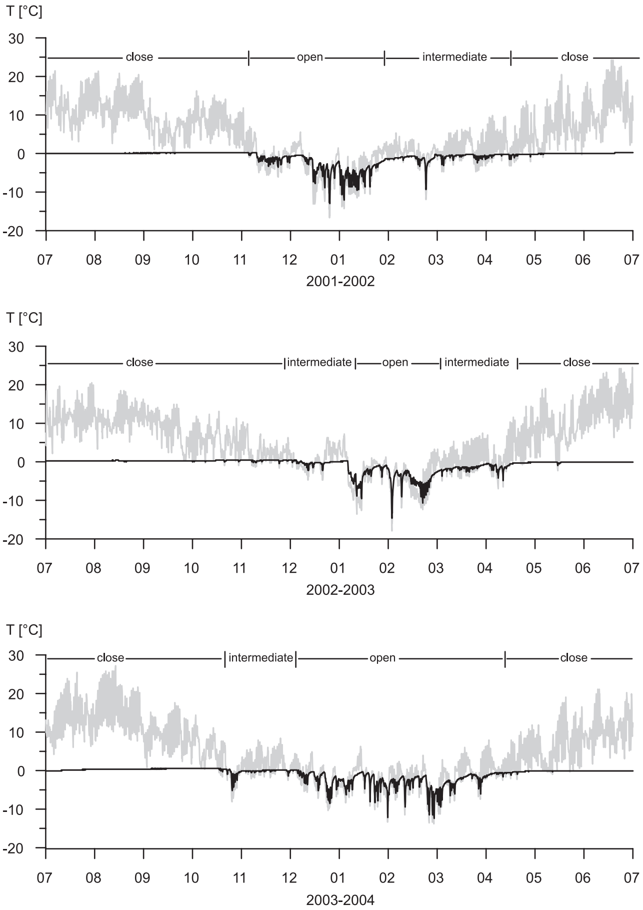


Fig. 32 Air temperature recorded at Monlési study-site between 07.2001 and 07.2004. Recorded data highlight an open period ranging between November and April which is characterized by a good correlation between the cave air temperature (black) and the negative exterior air temperature (grey).

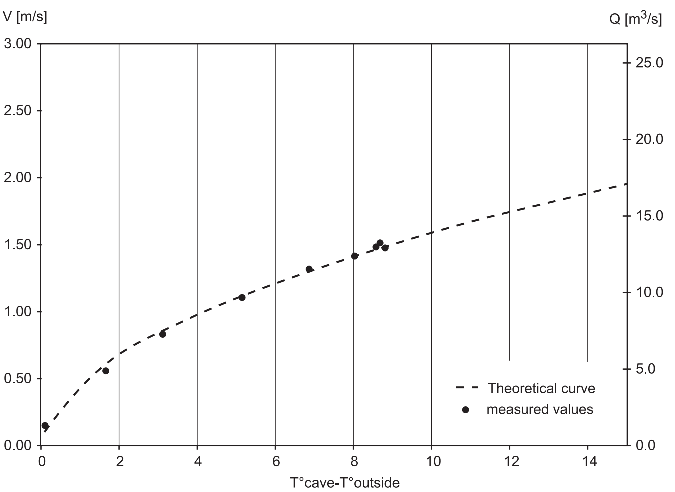


Fig. 33 Air flow measured at the main entrance of Monlési ice cave. Measured values fit well with the theoretical curve.

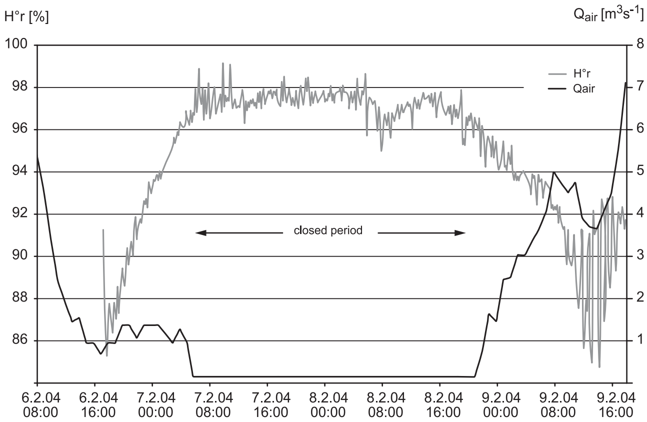


Fig. 34 Air flow and related cave air humidity inside Monlési ice cave. Ventilation episodes transport non-saturated air into the cave and therefore enable sublimation of cave ice.

6.2.2 Sensible heat stored in the ice filling

The temperature distribution within the ice filling deserves attention both for the role played in the energy balance of the ice cave as well as its relation with potential paleoclimatic records. Figure (35) illustrates calibrated temperature data recorded from November 2002 to October 2003 at depths of -1, -2.5, -4, -5 and -8 meters. Seasonal temperature oscillations are observed within the entire ice filling. Data recorded at -8 m suggest a temperate ice filling affected from all sides by melting processes.

Modelled temperature distribution was compared to measured values assuming a homogeneous ice body (i.e. constant heat diffusivity of $1*10^{-6}$ m^2s^{-1}) with no internal heat generation. Considering one-dimensional heat conduction, the temperature at a depth z is given by the diffusion equation:

$$\frac{\partial T}{\partial t} = a\frac{\partial^2 T}{\partial z^2} \qquad (6.1)$$

where T: Temperature [K]; t: time [s]; *a*: thermal diffusivity [m^2s^{-1}]; z: depth [m].

Equation (6.1) was solved numerically using a finite difference scheme (e.g. Mitchell & Griffiths, 1987). Convective heat transfers were disregarded and the temperature measured at 1 m depth was used as the input time-series. As illustrated by Figure (36), modelled values fit perfectly with data measured by the thermistor located at -2.5 m (r^2: 0.99). Since two-dimensional modelling improved the accuracy of the computed temperature distribution at greater depths, the uncertainties were mainly attributed to the geometry of the ice filling (e.g. Part II, p. 98). However, a significant discrepancy with measured values is observed with the thermistor at 5 m depth. This discrepancy was attributed to the presence of a major discontinuity allowing water drainage within the ice body. Such an interpretation is consistent with observations performed during the steam drilling. Nevertheless, on a large scale the ice body is quite homogeneous along the entire thickness.

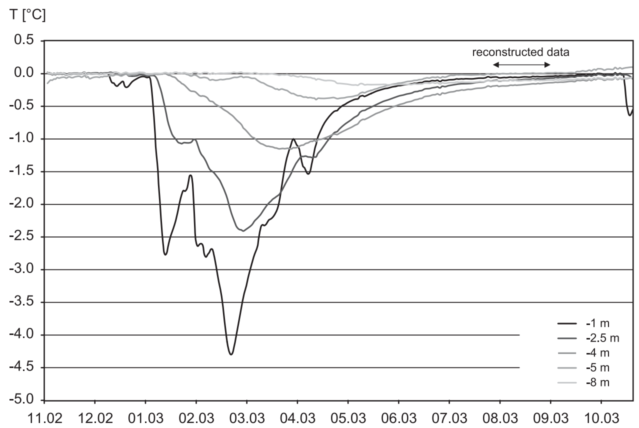


Fig. 35 Daily mean temperature recorded at different depths in the ice filling of Monlesi cave. Although seasonal fluctuations are observed over the entire filling, data suggest a temperate ice body.

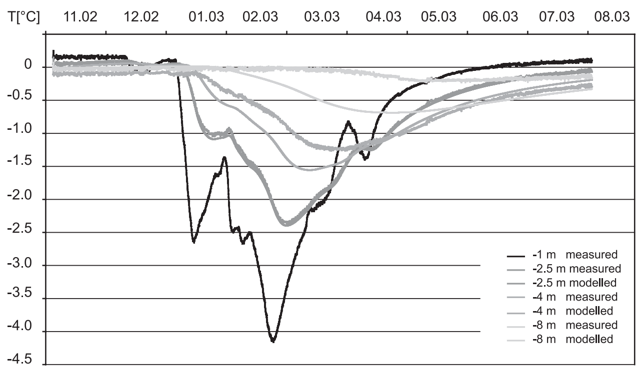


Fig. 36 Comparison between the recorded ice temperature and the modelled heat diffusion. Data suggest a 1D thermal diffusion within the first meter but lateral effects have to be considered further deep.

Considering that the recorded temperatures are representative of the entire ice filling (i.e. ~6000 m^3), the maximal sensible heat stored within the ice could be assessed at -11.6 GJ. The accuracy of this order of magnitude could be significantly increased by considering a 3D finite-element model, but uncertainties on the extension of the ice filling are too important yet.

Since the data demonstrated that the ice body almost reaches a thermal equilibrium in mid-August, it can be concluded that sensible heat transfer is not relevant for the annual energy balance of the ice cave. However, the cooling of the ice body tends to significantly reduce the duration of the melting period. During the 2002-2003 cycle, this difference was estimated to be about one month after the last freezing days. The duration is even longer if the heat storage within the surrounding limestone is considered.

6.2.3 The heat supplied by water infiltrations

Water discharges measured after a major precipitation event showed that most of the actual infiltration observed in Monlési ice cave is concentrated on one single inlet. Figure (37) illustrates the discharge monitored between November 2002 and October 2003. Maximal values measured during this time period reach nearly 15 lmin^{-1}. The analysis of flood events suggests a highly transmissive system with a strong control from external precipitation distribution. Although the data emphasizes a non-stationary relationship between precipitation and measured discharges, the hydrological balance of the system indicates a mean catchment area of 240 ± 25 m^2 for this specific inlet (e.g. Part II, p. 98). However, because of the dense vegetation in the catchment area, the mean specific infiltration observed between May and October was estimated at around 20 % of the actual precipitation.

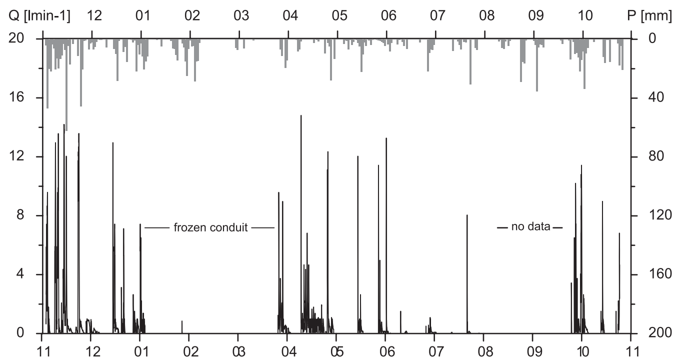


Fig. 37 Reconstructed water discharge at the main inlet in Monlesi ice cave. Major uncertainties remain on the reconstructed precipitation amounts, which tend to be systematically underestimated during thunderstorms. Reduced precipitation amounts, as mesured during the summer 2003, suggest that hot but dry periods are favourable for the consevation of cave ice.

According to the recorded data, the main water inlet remains frozen about 40 % of the time (3411 hours from 11.2002 to 11.2003). Water temperatures show a strong correlation with measured discharges. Maximal values are reached with high discharges, when heat exchange with the surrounding rock is minimal, and are assumed to correspond to the temperature at the system's boundary conditions within the limestone (i.e. exterior MAAT: 4.5 °C).

Although determining a reliable transfer function was not possible in a simple approach, empirical relations enabled reconstruction of the missing data. Accuracy on reconstructed discharges was estimated at ±30 % and modelled temperature curves fit measured values relatively well (r^2: 0.9).

The heat advected towards the system is defined as:

$$P = \frac{c_w \int_{t1}^{t2} Q_m \theta \, dt}{t_2 - t_1} \qquad (6.2)$$

Where P: mean power of inflowing water [W]; c_w: heat capacity of water [$Jkg^{-1}K^{-1}$]; Q_m: water flow mass rate [kgs^{-1}]; θ: temperature of water [°C]; t_1, t_2: time at instant 1 and 2 respectively.

Considering that the heat exchange with the cave is maximal, it was assumed that the outflowing water temperature is at 0°C. Hence, the thermal contribution of the main water inlet during the 2002-2003 annual cycle corresponds to about +3 GJ. Since it is known that this inlet drains about half of the water circulating through the cave, the total heat input to the system by water infiltration can be assessed at about +6 GJ during the observation period. In fact, the boundaries of the global system were defined by a constant rock temperature of 4.5°C suggesting further heat exchanges between the outflowing water and the surrounding limestone. Therefore, it is very likely that the thermal contribution of infiltration water has been slightly overestimated. But since it can be demonstrated that the contribution of water infiltrations to the energy balance of the cave is small, a more accurate model appears to be superfluous for our applications.

Thermal contribution of intrusive snow accumulations

According to daily measurements performed in La Chaux-de-Fonds (pers. comm. Travaux Publics), cumulated snow precipitation during the winter of 2002-2003 reached nearly 2 m thickness. Assuming that this value is representative of the Monlési site and that the mean snow density is about 300 kgm^{-3}, the snow mass accumulated at the base of the entrance shafts of Monlési ice cave (i.e. 140 m^2) was calculated at 84 tons. Thus, the latent energy supplied by intrusive snow accumulations is about -28 GJ during the 2002-2003 observation periods. Since a major uncertainty remains concerning the snow density, the final accuracy of this computed value is estimated at ± 30%.

Heat supply induced by ground heat fluxes

Rock temperatures measured in the Monlési ice cave show a good correlation with air temperature fluctuations, suggesting significant heat exchange between both systems. The almost constant temperature gradient observed in all four boreholes at the end of the summer season (Fig. 38) suggests a ground heat flux ranging between 1 and 2 Wm^{-2} under steady state conditions. This estimation is consistent with observations of basal melting issued from topometrical surveys (~10 cm $year^{-1}$). Assuming a constant cave temperature of 0°C, and considering a homogeneous limestone (λ: 2.2 $Wm^{-1}K^{-1}$) this heat flux enables one to assess the boundary condition of the rock system (i.e. constant temperature of 4.5°C) at a distance ranging between 5 and 10 m from the cave walls.

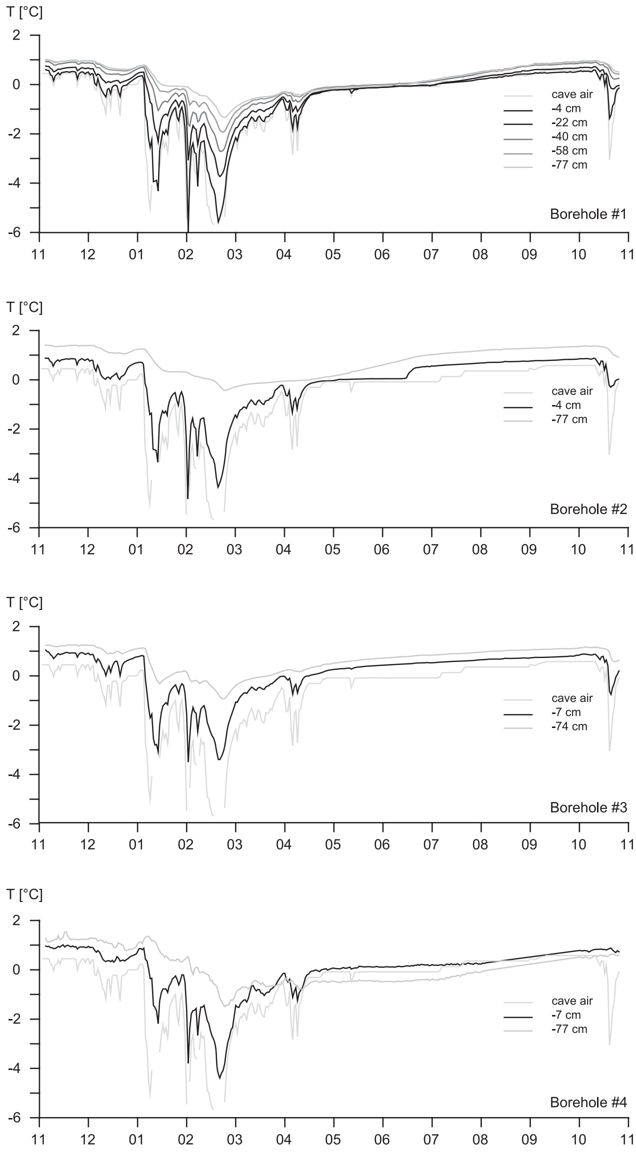


Fig. 38 Rock temperatures measured in four different boreholes of Monlési ice cave between 11.2002 and 10.2003.

Approximating the observed heat flux by a 1D heat conduction, the equation was solved in a finite-difference scheme. The reconstructed data provided clear evidence that the apparent heat diffusion within the limestone is significantly lower than expected and ranges between 0.3 and 0.5×10^{-6} m^2s^{-1} (instead of $a_{limestone}$: 1.13×10^{-6} m^2s^{-1}). Initially assuming the absence of water movement, this low value was interpreted as primarily related to latent-heat effects. The phase changes are well illustrated by the abrupt temperature change observed after a long stable period corresponding to freezing/thawing events (i.e. 0 °C; Fig. 39). However, since the modelled data does not entirely fit the measured values, the presence of advective fluxes had to be considered, suggesting a highly heterogeneous material.

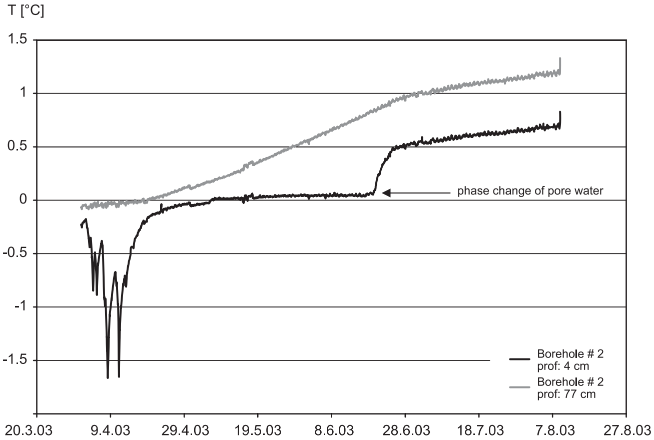


Fig. 39 Rock temperature recorded from April to July 2003 in Borehole #2. The thawing of the rock is evidenced by the phase change of pore water.

As a consequence, sensible heat storage was estimated empirically by discretizing the rock wall into smaller elements associated with the recorded data. Borehole #1 has been considered as representative of the rock temperature distribution within the cave. By assuming a rock-air exchange surface of about 1100 m^2, the upper limit of sensible heat stored during the 2002-2003 year is calculated at -9.2 GJ. The related uncertainty is estimated at ± 30%.
Since a mean constant ground heat flux of 1 Wm^{-2} is considered representative, the energy supplied by the surrounding massif is given by:

$$E_{rock} = \left(\int_{time} \varphi\, S\, dt\right) - E_{sens} \qquad (6.3)$$

where E_{rock}: energy issued from the ground heat flux [J]; φ: density of heat flux [Wm^{-2}]; S: exchange surface; E_{sens}: Sensible heat stored within the rock [J].

Measured rock temperatures enabled calculation of this energy supply at +117GJ for the 2002-2003 observation period.

6.2.4 The energy balance of the ice cave

The energy balance of the Monlési ice cave is defined as (cf. 3.3.1):

$$\Delta E_{air} + LE_{snow} + LE_{ice} = S + \Delta E_{water} + R \qquad (6.4)$$

Where ΔE_{air}: heat advected by air circulation; LE_{snow}: latent heat of intrusive snow; LE_{ice}: latent heat of ice; S: ground heat flux; ΔE_{water}: heat advected by water circulation; R= solar radiation.

Due to the cave morphology and the surrounding topography, direct solar radiation is almost never observed in the entrance pits. It was assumed that, due to the presence of dense vegetation at the shaft entrances, the diffuse radiative heat flux can be neglected. This simplification is further supported by the high albedo of the snow accumulated at the base of the entrance shafts. Hence, heat transfer observed during the 2002-2003 annual cycle enables quantification of the energy balance of Monlési ice cave as follows:

$$75\text{ GJ} + 28\text{ GJ} + LE_{ice} = 117\text{ GJ} + 6\text{ GJ} + 0\text{ GJ}$$

Assuming the hypothesis that the disequilibrium of the energy balance is completely compensated for by the melting or freezing of cave ice, the adjusting parameter LE_{ice} is estimated at an energy of some 20 GJ during the 2002-2003 annual cycle. This energy corresponds to the disappearance of an 8 cm thick ice layer which corresponds to observations performed on the fluctuation of the cave ice volume.

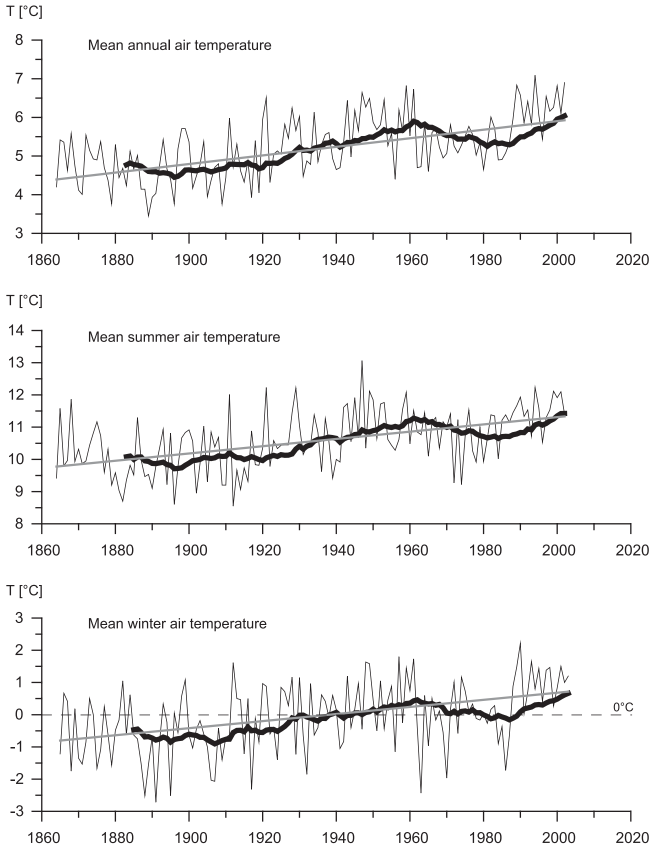


Fig. 40 Mean air temperature evolution measured in the Jura Mountains (Meteoswiss data). A twenty-year moving average (bold line) outlines a general trend of about 1 °C/100 years at the station of Chaumont (r^2: 0.77). This trend is more or less equally distributed between summer and winter semesters. But since the nineteen forties, mean winter temperatures, which were usually below 0 °C at this altitude, became mostly warmer than the freezing point.

6.3 The effect of climate forcing on the mass balance of cave ice

6.3.1 Observed variations in the regional climate

The analyses of monthly air temperature time-series, reconstructed and homogenized back to 1865 by MeteoSwiss for the station at Chaumont (6°59'16.3''E/ 47°3'4.7''N/ 1073 m a.s.l.), illustrated a general positive temperature trend of about 1°C/100 years (r^2 = 0.77). Although major regional disparities were observed, this trend is consistent with other recorded data (IPCC, 2001).

The tendency is more or less equally distributed between summer and winter semesters (Fig. 40). Regarding ice caves, a significant change occurred in the 1940s, when mean winter temperatures, which were mostly below 0°C at this altitude, rose above the freezing point. After this trend slightly reversed between 1960 and 1988, winters became clearly milder during the last decade of the twentieth century.

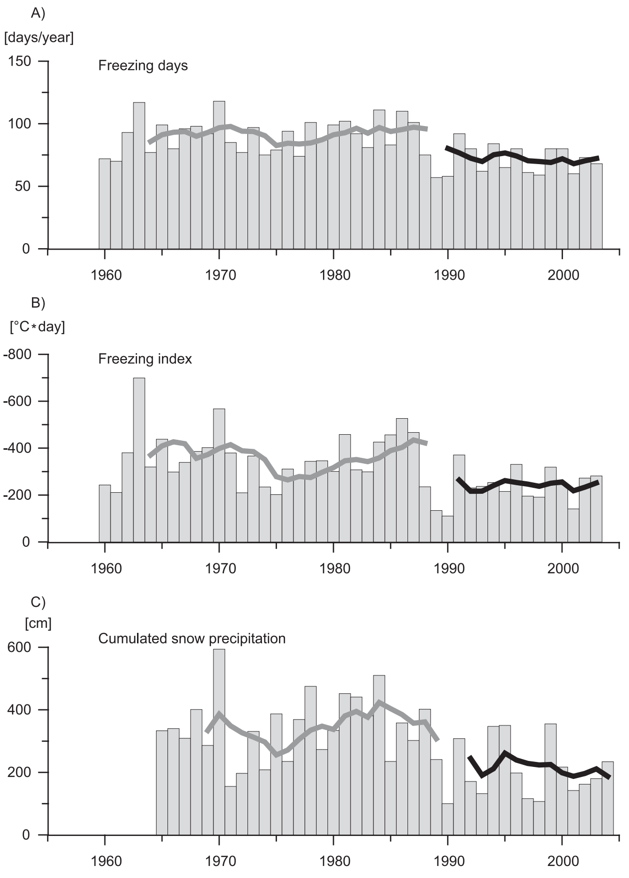


Fig. 41 Time series of winter climatic parameters in the Jura Mountains. A five-year moving average underlines the major change observed at the end of the nineteen eighties (bold lines). A) Number of freezing days (Daily MAT < 0°C); B) Freezing index [°C * day] and C) Cumulated snow precipitations [cm]. Daily records, provided for the station of Chaumont by Meteoswiss since 1959, demonstrate a significant change in winter climate since 1989: milder temperatures are observed simultaneously with decrease of snow accumulation.

Consequently, and especially since 1989, freezing days (number of days where Daily Mean Air Temperature - DMAT - is below 0°C) became less frequent and the annual freezing index («FI»: integer of DMAT<0°C) significantly decreased. This trend is well supported by Figure 41 a & b which outlines a reduction of the FI by about 40% (i.e. ~130°C*day) in the Jura Mountains.

Although long time-series do not show any significant variation in precipitation regimes during the twentieth century (Bader, 2002), low-altitude snow precipitations decreased to the concurrent temperature increase.

Since both winter cooling and the presence of water are necessary for the crystallization of new cave ice, the changing climatic context leads to a disequilibrium of the annual energy balance for low-altitude ice caves. This fact is well evidenced by the measurements performed during the 2002-2003 annual cycle and the consequences for the secular mass balance of cave ice were analysed further.

6.3.2 Reconstructed cave ice mass balances

Owing to the reduced heat exchange with the external atmosphere during the summer season, seasonal melting of cave ice accumulations is mostly controlled by water infiltration (e.g. thunderstorms). However, the thermal contribution of water infiltration remains reduced with regard to the overall energy balance of an ice cave. Therefore, it can be concluded that, contrary to alpine glaciers, summer climatic conditions do not play any significant role in the mass balance of cave ice. This conclusion is well supported by observations made in numerous ice caves in the Jura Mountains (Luetscher et al., 2005; Part II, p. 65).

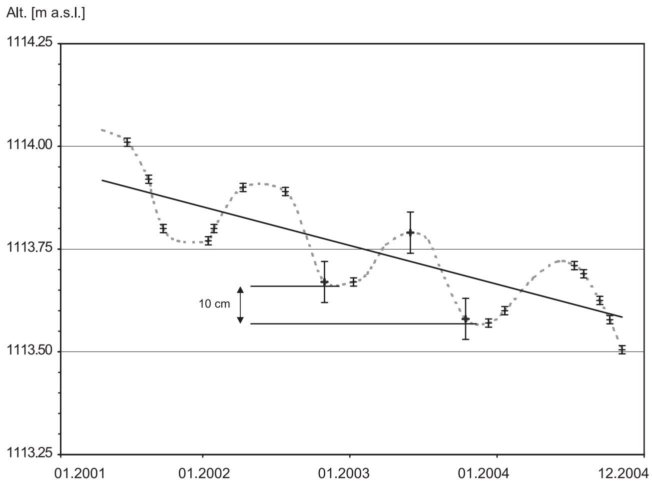


Fig. 42 Ice fluctuations measured in Monlesi ice cave between 2001 and 2004. Vertical bars reflect the accuracy of measurements. The thickness of the ice body decreased of about 10 cm during the 2002-2003 annual cycle. Although a negative trend of the ice mass balance can be observed, an exceptionally hot period in the summer of 2003 had no significant impact on the ice volume.

Figure 42 illustrates ice fluctuations measured in the Monlési ice cave over a three-year observation period. The data reflects the level of the ice surface at a station located on top of the ice filling. The marked seasonal variations confirm that the crystallization of cave ice is still an active process. Although a negative trend of the annual mass balance is observed, no significant increase of melting rates was mesured during the summer of 2003 which was characterized by exceptionally elevated temperatures in Western Europe (e.g. Beniston, 2004).
Even if observations are scarce and mainly recent, a reconstruction of long-term fluctuations was attempted based on former topographic surveys and photographic documents. Figure 43 shows a reconstruction of the evolution of ice volumes in five of the documented caves. Although major uncertainties remain on the geometry of each filling, relative fluctuations could be set in evidence. This compilation confirmed a general decrease of ice volumes during the twentieth century in most ice caves in the Jura Mountains. However, the trend is all but consistent over the entire observation period. Mass balances were probably close to the equilibrium up to the end of the 1980's. After that, a noticeable decrease of ice volumes is observed over the last fifteen years. This trend correlates well with the observed increase in winter temperatures, subsequently reducing the potential freezing of infiltration water. Hence, massive congelation-ice speleothems like those observed in St-Livres ice cave almost completely disappeared (Fig. 44a & b). This negative trend of cave ice mass balances, however, is evidenced in vertical shafts where reduced winter snow accumulations are no longer able to compensate for the annual melting (for instance, Gouffre de Bellevue, Fig. 45). Consequently, many of the smallest cave ice deposits have completely disappeared.

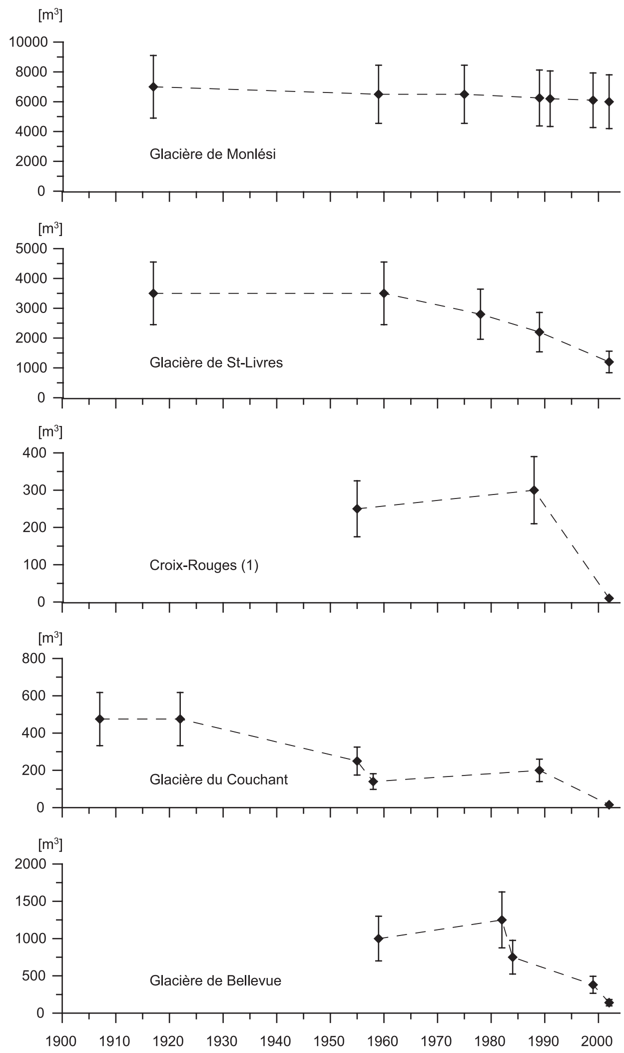


Fig. 43 Reconstructed cave ice volume fluctuations on selected sites in the Jura Mountains. Mass balance estimates outline a strong decrease of cave ice during the last decade. Error bars reflect the estimated inaccuracy (30%) attributed to measurements and seasonal variations of ice volume.

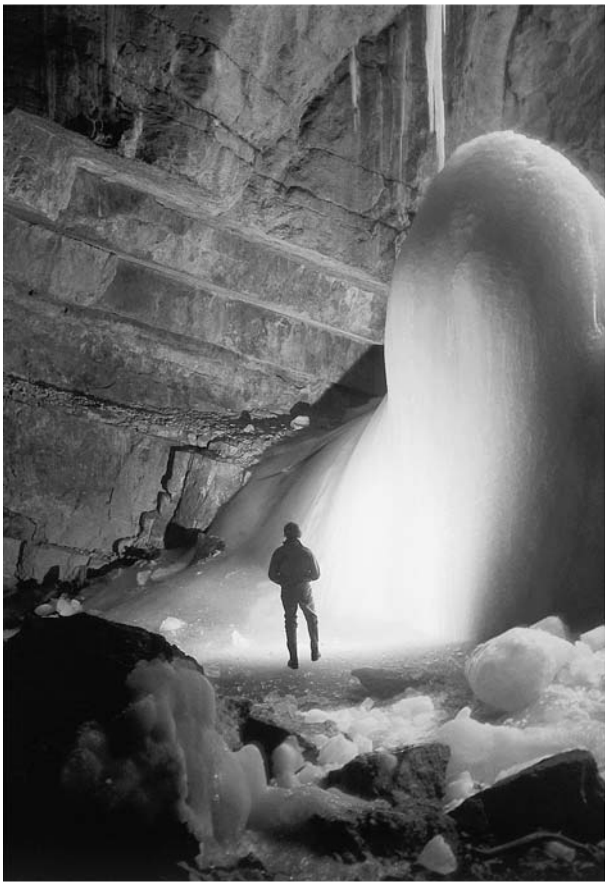


Fig. 44a Comparison of two pictures from St-Livres ice cave taken in 1978 (Fig. 44a) and 2002 (Fig. 44b) respectively. The negative cave ice mass balance could be attributed to milder winter temperatures and reduced snow precipitations observed since 1989.

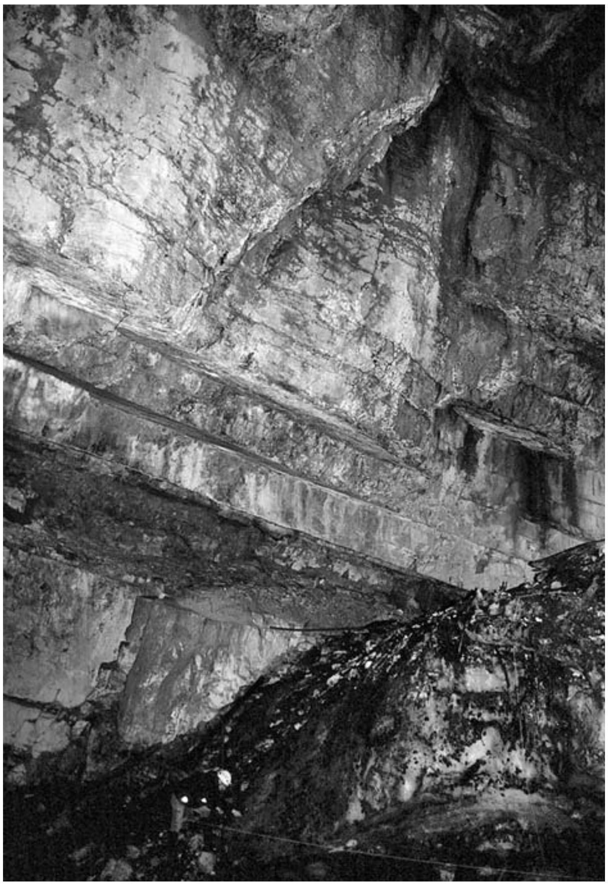


Fig. 44b

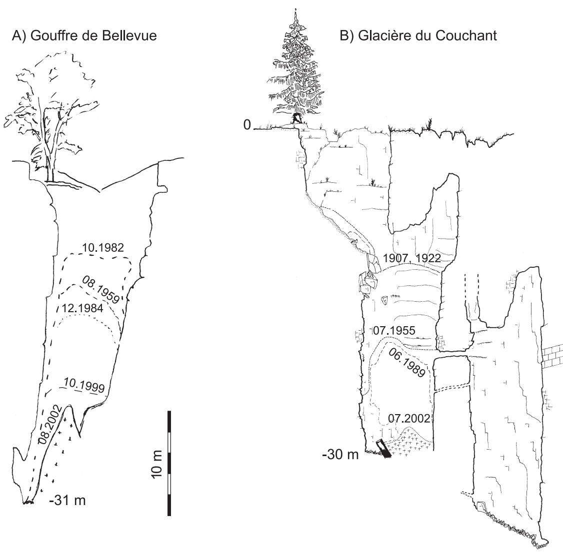


Fig. 45 Firn extension in two static ice caves. Both vertical cross sections show the evolution of the observed ice filling during the last decades of the twentieth century. Mass balance seems to have been quite stable until the end of the eighties but a strong negative trend is observed since 1989.

6.4 Periglacial records in cave sediments

Although subsurface ice accumulations completely disappeared from many caves of the Jura Mountains, speleological observations nevertheless indicate that periglacial records could still be well preserved in cave sediments. As suggested by Mihevc (2003), cryoclasts remain the most common features observed in caves. Actually, sharp, irregular, blocky fragments characteristic of ice wedging are found in almost every ice cave of the Jura Mountains (Fig. 46). Mainly observed at cave entrances, such patterns are sometimes described as deep as 100 m below the surface (for instance in the Glacière de Druchaux, Berolles, Vaud). Thus, Pancza (1992) attributed cryogenic processes with a major role in the morphology of ice caves.
However, cyclic freezing and thawing of cave sediments also leads to cryoturbatic ground movements. Figure 47 illustrates some of the sorted patterns found in the Glacière des Baumes (Les Verrières, Neuchâtel). Since stone circles, polygons, stripes and clay hummocks are sometimes observed deep within the cave (e.g. Mihevc, 2003), fossil patterns constitute a good indicator of a colder climatic context. In fact, autochthonous cryoclastic cave sediments have been described in deep alpine caves (for instance Hölloch, CH) where they were taken as evidences of freezing temperatures during the Pleistocene glacial maxima (Schmid, 1958).

Fig. 46 Cryoclastic sediments observed in Monlesi ice cave. Observations attributed the ice wedging to the action of hoarfrost. Photo M. Luetscher

Ground cryoturbation features and/or indications of former ice fillings are described in caves of the Jura Mountains at altitudes as low as 525 m a.s.l. Owing to the reduced erosion occurring in fossil caves, such patterns are preserved over several thousand years. In addition to traditional investigations, subsurface sediments in low-altitude caves could be good archives of recent winter climate fluctuations.
Furthermore, such evidence of a periglacial context val-

idates the hypothesis that the Equilibrium Line Altitude (ELA) of ice caves moved up several hundred meters when compared to the colder periods of the Holocene.

Fig. 47 Sorted stone patterns (scale=30 cm) observed in the glacière des Baulmes (Verrières / NE). The disappearance of numerous low altitude ice fillings lead to distinct periglacial signatures in several caves. These can be used for investigating winter paleoclimate conditions. Photo M. Luetscher

7 | Discussion

Field knowledge constitutes the indispensable background for the understanding of any environmental system. By documenting systematic observations carried out over a century, speleologists provided the basic elements for a process-based understanding of ice caves. The synthesis of historical data and its comparison with present-day observations enables a qualitative description of the main processes at the origin of ice caves in the world and especially in the Jura Mountains. However, quantitative estimations are necessary for the assessment of the relative importance of the identified processes. Only a process-based description makes it possible to interpret subsurface ice patterns in their environmental context.

7.1 Ice caves as an environmental marker

Starting the study of ice caves from the basic physical laws that govern subsurface heat exchange constitutes an effective tool for understanding the evolution of ice caves under changing boundary conditions (for instance external climatic conditions). Since the physics of the investigated system remains invariable through time, the chosen approach is expected to apply to all considered time periods. Thus, if the related equations could be solved perfectly, high confidence in predictions of the future evolution of ice caves could be achieved. Owing to the inherent uncertainty, calibration of the model is necessary using detailed field measurements under the present-day climate context. The approach proposed for the Monlési ice cave relies on the energy balance of the investigated system (Fig. 48).

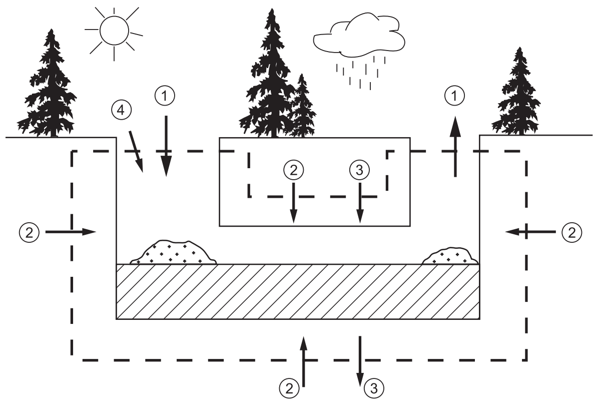


Fig. 48 Process-based model of heat exchanges in an ice cave. Heat exchanges observed at system boundaries comprise (1) Forced convection; (2) Conduction through the surrounding limestone; (3) advection by water circulations; (4) solar radiation.

Heat exchange included in this model comprises: (1) forced convection; (2) advection by water circulation or snowdrift; (3) conduction through the surrounding limestone; and (4) thermal radiation. The energy balance of this system corresponds to the decrease/increase of the ice volume as long as water is available for the accumulation of ice. Depending on the predominant process (forced convection or advection by snowdrift), our study suggests that alpine ice caves could be classified into nine different types (cf. 2.5.5 and Luetscher & Jeannin, in press a; Part II p. 59). Beyond these types, three principal factors determine the region of conditions where the presence of mid-latitude/low-altitude cave ice is possible:

1. the localization of the cave;
2. the morphology of the cave;
3. the local environmental context.

All of these three factors will be discussed in the following sections by analysing the heat transfers sketched on Figure 48. The aim of this discussion is to show to what extent the conceptual model developed for the Monlési study site can be generalized and applied to other ice cave systems.

7.1.1 Forced convection

Field measurements completed in the Monlési ice cave show that cold air circulation during the winter season constitutes the major process of heat exchange between an ice cave and the external atmosphere. This observation is consistent with early investigations carried out in the Austrian Alps (e.g. Bock, 1913; Saar, 1956). In fact, the present study demonstrates that forced convection takes place even if the altitude difference between the cave entrances is negligible. Although it contradicts the model proposed by Lismonde (2001), our field measurements demonstrate that the air flows always in the same direction. This observation is attributed to the contrasting dimensions of the various entrance shafts. In agreement with Bock's (1913) calculations, latent heat exchanges associated with winter air circulation are assumed to be a major factor in the energy balance of an ice cave. As suggested by Saar (1956), field data confirms that phase changes related to sublimation contribute further to the cooling of the cave during the open period. Measurements from the Monlési ice cave show that this latent-heat exchange roughly equals the sensible heat flux induced by the air circulation. These conclusions suggest that heat exchanges are lowered during humid weather conditions, which are frequently associated with small temperature differences between the cave air and the external atmosphere (i.e. reduced airflows). Therefore, continental climate regimes favour the formation of congelation cave ice. However, because of the slow kinetics of heat exchange between flowing air and cave walls or ice, the efficiency of heat transfer decreases with increasing airflow. Therefore, heat exchange is assumed to be maximal during long periods with small temperature differences between the cave air and the external atmosphere rather than during brief and very cold events. But, converse to Stettler's (1971) conclusion, evaporation appears to be insignificant during the closed period. Therefore, this process is not relevant for the conservation of cave ice during the summer season.

In our model, an increase in altitude difference between the cave entrances would turn the system into a real dynamic ice cave. The resulting improved chimney effect would make the cooling of the ice cave even more efficient during the winter season. However, as soon as the external temperature rises above the cave temperature, this airflow would be reversed supplying a significant amount of heat to the system. Ice crystallization during the winter season must therefore be sufficient to maintain the equilibrated energy balance of the ice cave. By considering seasonal and daily exterior temperature fluctuations, it is estimated that perennial conservation of congelation cave ice (stable mass balance) is limited, in Switzerland, to altitudes above about 1600 m a.s.l. Although this value is closely related to local climatic parameters (mainly related to the topographical context), the absence of speleological prerequisites signifies that dynamic ice caves are rarely possible in the Jura Mountains, where summits culminate at 1700 m a.s.l.

The ventilation observed in a single downward-sloping cave which leads to an obstructed conduit is a particular example of a chimney effect. This statement is consistent with Bock's (1913, p. 118) conclusion and suggests that the airflow observed in static caves depends on the altitude difference between the entrance and the deepest point of the cave. Furthermore, since heat exchange is enhanced by large exchange surfaces, the presence of static ice caves in mid-latitude/low-altitude environments is often associated with large cave volumes. In summer, the conservation of perennial cave ice is favoured by limited heat exchange between the cave air and the external atmosphere. This characteristic is attributed to the well-described thermal trapping induced by the density difference between the internal and external air masses (e.g. Thury, 1861). Owing to the insignificant altitude difference observed between the various cave entrances, the Monlési ice cave has such a characteristic. As already described by Browne (1865), there is a small oscillating airflow between the different entrance shafts during the closed period. Measurements demonstrate that the heat supplied by this airflow is rather insignificant as compared to the overall energy balance. Nevertheless, the conclusions issued from this study are in contradiction with Stettler & Monard (1960), claiming that the 0 °C cave-air temperature observed during the summer season was mainly related to the crystallization of hoarfrost.

However, the specific morphology of the Monlési ice cave (i.e. several cave entrances located at the same altitude) represents almost an optimal condition for an efficient cooling of the system. Sensible heat exchange induced by cold-air circulation during the winter season is enhanced by the latent-heat transfer produced by sublimation. Furthermore, the cold-air trapping observed during the summer season significantly reduces the heat exchange with the external atmosphere. Thus, combining both features leads to a major temperature anomaly within the karst system. The heat transfer is significantly enhanced by the freezing of infiltration water.

7.1.2 Heat advected by water circulation and snow drift

Heat advected by water infiltration is closely related to the discharge rate at the inlet, which is controlled by the actual infiltration and the drainage pattern of the karst system. Consequently, the presence of cave ice is restricted to locations where infiltration is diffuse rather than concentrated. Very high infiltration rates are not favourable for ice caves. Trimmel (1969) suggested that a reduction of the actual infiltration induced by increased evapo-transpiration related to a dense vegetation cover at the surface could favour the conservation of cave ice. Although this point of view is consistent with a hydrological balance, our study demonstrated that, in the Monlési ice cave, the heat supplied by water infiltration is rather insignificant. Nevertheless, because the efficiency of heat transfer decreases with increasing water discharge, ice crystallization occurs preferentially at smaller water inlets. Field observations confirm that ice is formed when cold percolation water (for instance snow meltwater or rain) reaches a cave with temperatures still below 0°C. Thus, positive cave ice mass balances are observed mostly in spring, when warm days allowing snowmelt are followed by cold nights. However, mid-winter warm spells favour the crystallization of high amounts of congelation ice. Therefore, the results further support Rachlewicz & Szczucinski's (2004) suggestion that long mild winters are more favourable for ice growth than very cold winters. Contrary to water infiltration, snowdrift at the cave entrance is a major factor in the energy balance of an ice cave. This contribution is closely related to the dimension(s) of the entrance shaft(s) and to winter precipitation rates. Thus, major subsurface snow accumulations are mostly found at the bottom of broad entrances located in mountainous regions with a humid temperate climate.

7.1.3 Conduction through the surrounding limestone

The conceptual model of temperature distribution in karst systems proposed by Luetscher & Jeannin (2004; Part II, p. 52) suggests that the lower and lateral boundary conditions of the ice-cave system is defined by a constant rock temperature corresponding to the external Mean Annual Air Temperature (MAAT). Therefore, as suggested by Saar (1956), mid-latitude/low-altitude ice caves are major thermal anomalies within local karst systems, leading to significant temperature gradients in the surrounding limestone.

Rock temperature measurements achieved in the Monlési ice cave demonstrated that the related ground heat fluxes constitute the most significant energy supply to the ice-cave system. Although this conclusion is validated by a monitoring of high melt rates at the rock-ice interface, it contradicts data from many other study sites where basal melting rates were considered to be rather insignificant (for instance, Rachlewicz & Szczucinski, 2004). Several hypotheses are proposed to explain the difference in observed ground heat fluxes. Conductive heat exchange with the surrounding rocks depends on the temperature gradient measured between the cave wall and the boundary condition. Assuming an upper limit of the cave temperature of 0°C, this gradient depends on the distance to the depth of constant temperature as the boundary condition and on the MAAT. As the latter decreases linearly with altitude, the related ground heat flux is expected to be inversely proportional to the altitude of the ice cave. Borehole temperature logging demonstrates that the limestone can hardly be considered as homogeneous on a metric scale. As confirmed by data from the Monlési ice cave, air and water flows control the temperature distribution within the rock walls. Therefore, local variations in the drainage network around the ice cave easily explain significant fluctuations in the distance between the cave and the constant temperature boundary condition. In the same way, the presence of porous materials (for instance a scree slope) between the principal ice mass and the bottom of the cave could allow the local cold air circulation to drain the ground heat flux from the underlying karst system. Cold-air traps are frequently associated with large cave sediment deposits at the cave bottom. This indicates that because of the high porosity of the sediment, conductive heat transfers are often low enough to prevent any significant melt rate at the rock-ice interface.

The high thermal inertia of the rock system makes the ground heat flux almost constant through decadal timescales. Field observations performed in Monlési ice cave validated this interpretation and verified the absence of any exceptional ice decrease during the 2003 summer heat wave. However, long-term temperature fluctuations will finally modify this boundary condition. It can be postulated that warming climate causes an increase in ground heat fluxes according to the long-term trend of MAAT. Observations from the Monlési ice cave suggest that, depending on the annual crystallization rate, an equilibrated cave ice mass balance could theoretically be observed with a mean ground heat flux of about 1.2 Wm^{-2}.

This implies a constant temperature boundary condition of 5.4°C, which corresponds to a mean annual air temperature of nearly 1°C higher than the 2002-2003 value. This calculation corresponds with estimations provided by Saar (1956). However, as suggested by photographic comparisons, the yearly cave ice mass balance was mostly negative since 1989 owing to low crystallization rates attributed to mild winter temperatures and reduced snow precipitation.

7.1.4 Radiation

The energy balance of the studied system relies on the assumption that thermal radiation can be neglected. This assumption is supported by observations performed in the Monlési ice cave which confirm the absence of direct radiation in the entrance shaft areas, partly attributed to the dense vegetation cover. This scenario is consistent with Vincent (1974) who considered the external vegetation as a secondary factor controlling the ice-cave distribution. Solar radiation is closely related to the topographic location of the study site, but the potential direct radiation is also a function of the diameter and depth of the entrance shafts. Thus, heat exchange with the external atmosphere is significantly reduced in cases where deep and/or narrow shafts shadow the cave entrance. Consistent with spatial modelling of permafrost distribution patterns (Hoelzle et al., 1993), the absence of direct solar radiation is considered a prerequisite for the presence of mid-latitude/low-altitude ice caves.

7.1.5 Generalizing the conceptual model at the origin of cave ice

Although several morphological parameters have to be considered in the energy balance of an ice cave (for instance, dimensions of the entrance shafts; cave volume; exchange surface with the cave air), the conceptual model of heat transfers developed for the Monlési ice cave is reliable for a process-based understanding of alpine ice caves. It is concluded that historical distinctions between static and dynamic ice caves concern particular cases of subsurface ice patterns. Thus, the selection of a stato-dynamic cave like the Monlési ice cave as a principal study site appears to be justified for a general understanding of mid-latitude/low-altitude cave ice patterns.

Besides some uncertainties inherent to the approximation of the cave and ice geometry, several simplifications were introduced to quantify heat-transfer processes in the Monlési ice cave. Therefore, the model could be significantly improved by further studies dedicated to local heat transfer. One of the most relevant aspects would certainly be a better quantification of latent heat exchange induced by sublimation. For this purpose, continuous measurements of air humidity at the system's boundaries would be necessary. However, such measurements would be confronted with significant technical constraints related to power supply. Other studies are needed to better estimate the ground heat fluxes. Special attention should be given to the spatial distribution of the rock temperatures around the cave. This could be achieved easily with infrared surveys of the cave walls. However, further boreholes should be drilled to improve the understanding of the system's boundary conditions. Unfortunately, data issued from quantification of observed heat transfers in the Monlési ice cave are confined to this particular cave and cannot be easily extrapolated to other study sites. Thus, a similar investigation at selected comparison sites is recommended. However, the main conclusions from the present investigation are validated by the general decrease of cave ice volumes observed during the last fifteen years in the Jura Mountains. This well documented trend is consistent with meteorological observations documenting the occurence of milder winter temperatures and reduced snow accumulations since 1989.

Results from the Monlési study site suggest that the ideal «mid-latitude/low-altitude» ice cave acts as a thermal trap during the summer season, and thus has one or several entrances which are located at the same altitude and leading to a clogged conduit. Forced convection during the winter season is favoured by the presence of several shafts and causes an efficient cooling of the system. Since the resulting airflow is closely related to external freezing air temperatures, caves at lower altitude (i.e. high MAAT) must have larger exchange surfaces. Major snowdrifts at the cave entrances favour a perennial conservation of cave ice. Deep entrance shafts shaded by vegetation best conserve these snow accumulations. Although seasonal hoarfrost patterns are frequently observed near ice-cave entrances, their sensitivity to air drafts prevents the formation of perennial deposits. Therefore, perennial hoarfrost is confined to continuous permafrost regions (i.e. high-latitude and/or high altitudes).

7.2 Low-altitude cave ice in a changing climate context

Because the presence of cave ice is controlled by specific climatic conditions (i.e. winter precipitation, number of freezing days, MAAT), ice caves are reliable environmental markers. Based on the model developed for the Monlési ice cave, a qualitative interpretation of the effects of climate change was attempted for two distinct scenarios, the one corresponding to a colder climatic context («Little ice age») and the other corresponding to warmer context («Global warming»).

7.2.1 Scenario I: «Little ice age»

Instrumental records and climatic reconstructions demonstrate the existence of a relatively cold period from the seventeenth through the nineteenth centuries in the Northern Hemisphere. This «Little Ice Age» period was characterized by a lower MAAT than today (<1 °C) with increased cold days and frost days over nearly all land areas (IPCC, 2001). Thus, snow precipitation at low altitudes was generally higher than observed today. Under such a climatic context, the Monlési ice cave is believed to maintain a positive ice mass balance. The cooling of the system by ventilation is improved by the increased number of freezing days. This effect is further increased by larger snow accumulation at the base of the entrance pits. Since the amount of percolating meltwater is reduced during the winter season, however, heat transfer is not as efficient as could be expected.

Although a positive cave ice mass balance is expected, the extension of the ice volume is limited by the air ventilation itself. Since the airflow cooling the system is closely related to the diameter of the conduit, it is significantly reduced if the ice volume fills most of the available space within the cave. At the same time, heat exchange with the surrounding rock is continuous, though slightly reduced because of the lower temperature at the boundary condition. Hence, although the cave ice volume would be larger than observed today, the system would remain controlled by the same processes as today. With the MAAT dropping below 0 °C, heat exchange with the surrounding limestone becomes insignificant in the overall energy balance, and ice melting at the cave bottom is strongly reduced. However, as suggested by Ciry (1962; p. 31), the permafrost observed below the active layer would significantly reduce the actual infiltration within the cave. Hence, active congelation-ice deposits are insignificant under such conditions. Nevertheless, the evaporation / condensation processes induce significant mass.

Although the conclusions developed for the Monlési ice cave are valid for numerous ice caves characterized by (stato-) dynamic ventilation, static ice caves with firn accumulations constitute a slightly different case. In fact, increased snow precipitation leads to major perennial firn accumulations in low-altitude ice caves. Hence, numerous «snow pits» are expected in traditional cold air traps. Their conservation is favoured by longer freezing periods.

7.2.2 Scenario II: «Global warming»

All scenarios issued from recent climate modelling tend to agree that climate forcing will lead to a temperature raise of at least 1.4 °C by 2100 (IPCC, 2001). Temperature increases are projected to be greater during the winter season than during the summer season and it is likely that fewer cold and frost days will be observed over nearly all land areas. Hence, these models suggest a significant decrease of snow precipitation at low altitudes.

The relative importance of the parameters involved in the energy balance of Monlési ice cave shows that ventilation of the system would be significantly reduced in such a climatic context. However, the efficiency of heat transfer increases if the freezing of percolating meltwater enables the formation of cave ice. As crystallization becomes almost impossible with elevated discharge rates (i.e. major precipitation events), the optimal equilibrium state is reached with the daily melting of the external snow cover followed by freezing nights. Since the frequency of such oscillating freeze/thaw cycles increases significantly with a warming climate, an unexpected positive cave ice mass balance may be observed if external air temperatures are fluctuating around 0 °C. Hence, contrary to the Northern Hemisphere snow cover which is projected to decrease further during the twenty-first century, some low-altitude ice caves could present a stable decadal mass balance under favourable climatic conditions. At this point it is not clear whether the ice volume in the Monlési ice cave would remain stable or decrease in the future. Based on our data, several modelling scenarios would provide more precise predictions. Although the Monlési ice cave and other similar stato-dynamic ice caves could maintain ice in a warmer climate, subsurface low-altitude perennial firn accumulations will probably continue their widespread retreat observed over the last two decades. A general increase in the lower limit of ice-cave occurence may be observed because the equilibrium state of such systems will no longer be reached. Hence, ice caves are associated with an equilibrium line altitude (ELA) dependent on the cave morphology and external winter conditions. Oscillations of this ELA could lead to the complete melting of several subsurface ice bodies. Since the presence of former ice fillings is recognized in cave sediments, such evidence provides useful information on past winter climatic contexts.

7.3 Paleoclimatic outlooks

7.3.1 Periglacial records in cave sediments

This study has shown that evidence of periglacial contexts is found in several caves of the Jura Mountains where perennial cave ice is observed. The reduced erosion/ alteration rates observed in many caves also demonstrates that periglacial records might be well preserved in cave sediments, among which cryoclasts remain the most common features (e.g. Kempe & Rosendahl, 2003). Sharp, irregular, blocky fragments are characteristic of ice wedging near cave entrances (White & White, 2000) and were sometimes described in deep alpine caves where they were taken as evidence for very deep freezing temperatures during the Pleistocene glacial maxima (Schmid 1958). With freezing and thawing cycles, cave sediments are subject to significant ground cryoturbations

leading to sorted patterns such as stone circles, polygons, stripes and clay hummocks (Mihevc, 2003).
As demonstrated by Zak et al. (2004), dating of former periglacial environments is possible from cryogenic cave calcite. Furthermore, Kempe & Rosendahl (2003) observed that small movements of cave ice could explain widespread damages observed on speleothems. Because such patterns might be preserved over thousands of years, our observations provided further evidence that cave sediments could be a good marker of Holocene winter climate fluctuations.

7.3.2 The accessible time period for glaciological investigations

Multi-parametric dating attempts of cave ice in two selected study sites of the Jura Mountains demonstrated that accessible time periods could vary significantly from one cave to the other. These differences are chiefly attributed to distinct mass turnover rates, which are controlled by ground heat fluxes at the system's boundaries (cf. 7.1.3). Since cave ice represents a valuable object for the reconstruction of winter climate evolution, explorative investigations were dedicated to the identification of potential proxies. Because remobilization of chemical signatures from percolating meltwater was expected, the focus was set on stable isotope data. Results suggested that the observed oscillating signature could be interpreted as the effect of successive fractionation processes induced by air circulation in contact with cave ice. Hence, high-resolution sampling combined with precise dating of each stratigraphical layer provides a valuable record of successive cold events during the winter season. In the Jura Mountains, such information is related to easterly winds supplying glacial air. Cave ice can be preserved over several centuries if not thousands of years. Located close to the source regions of atmospheric pollutions, mid-latitude/low-altitude cave ice represents an important archive of climate history in central Europe. Potential data from pre-industrial times could provide information of natural concentration levels for various atmospheric species, which are now influenced by human activities. Although remobilization of chemical components induced by percolating meltwater could significantly influence the quality of this archive, cave ice constitutes a suitable alternative to glacier records.

7.4 Recommendations for further investigations

Paleoclimatic interpretations of ice records require a perfect understanding of climatic processes involved in the cave ice deposition. The model of heat transfer in ice caves based on data from the Monlési study site is highly dependent on the detailed knowledge of the geometry of the thermal exchange surfaces. To increase the quality and general applicability of such models, special efforts should be devoted to the description of the 3D cave geometry, and especially to the ice volume. The shape of the ice bottom could be obtained by ground penetrating radar surveys using shielded antennas. A numerical model of evaporative processes along a karst conduit would be a useful tool for the interpretation of stable isotope analyses of cave ice. Since the cave air humidity increases along a conduit, the fractionation of oxygene isotopes in water is assumed to decrease with distance. Comprehensive empirical data could provide the background for the calibration of such a model. The main difficulty would be to obtain a precise and continuous record of the cave air humidity. This could be performed with a dew-point mirror and a high spatial resolution of temperature records, but issues with the power supply could be a problem. However, these results would constitute the indispensable data for a better investigation of heat transfer functions between the cave air and the surrounding rock.
The low mass-turnover rate observed in St-Livres ice cave prompts a better understanding of this ice body. Borehole temperature logging in the ice could prove the presence of a cold ice body suitable for further paleoclimatic investigations. Sequence stratigraphy from the ice front would help in the reconstruction of a complex sedimentary ice deposit. Looking at the organic layers enclosed within the ice could provide a link to external climatic fluctuations. Dendro-chronological methods constitute a reliable tool for a precise dating of different ice layers and thus, for identifying major time gaps in the stratigraphy. However, a better resolution is expected if the spruce-sample data analysis is compared to data collected from other tree species.
Future glaciological investigations of cave ice in the Jura Mountains will rely on the accessibility to reliable ice cores. As suggested by the present study, high-resolution dating can only be achieved using a multi-parametric analysis. Analyses of Tritium and Cesium-137 provide valuable results for the dating of recent cave ice deposits. Dating of secular ice layers may be possible under certain circumstances using the radiogenic decay of lead-210. Unfortunately, since major time gaps are expected in the stratigraphic sequence, counting of annual layers may not be feasible. This method could, however, be suitable for an age estimation of distinct ice layers between two absolute datings. Therefore, the most accurate method for older cave ice deposits still relies on the radiogenic decay of carbon-14. Encouraging results could be expected from the dating of cryogenic calcite.
Comparative data from other study sites would be highly valuable and would provide additional elements needed to estimate the suitability of ice caves as paleoenvironmental archives.

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PART II:

PROCESSES INICECAVES

Related Publications

A. Major journal papers
B. Further papers related to the thesis

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Temperature distribution in karst systems: the role of air and water fluxes

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ABSTRACT

A better understanding of heat fluxes and temperature distribution in continental rocks is of great importance for many engineering aspects (tunnelling, mining, geothermal research, etc.). This paper aims at providing a conceptual model of temperature distribution in karst environments which display thermal 'anomalies' as compared with other rocks. In temperate regions, water circulation is usually high enough to 'drain-out' completely the geothermal heat flux at the bottom of karst systems (phreatic zone). A theoretical approach based on temperature measurements carried out in deep caves and boreholes demonstrates, however, that air circulation can largely dominate water infiltration in the karst vadose zone, which can be as thick as 2000 m. Consequently, temperature gradients within this zone are similar to the lapse rate of humid air (~0.5 °C 100 m^{-1}). Yet, this value depends on the regional climatic context and might present some significant variations.

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Introduction

The temperature distribution in the subsurface is, in addition to its scientific interest, of great importance for various fields of practical applications such as for the construction of underground facilities, the use of groundwater resources or the development and exploitation of geothermal energy (Medici and Rybach, 1995). Owing to technical limitations, working conditions or sanitary considerations, elevated temperatures might frequently represent a source of problems for applications such as those mentioned above. Karst systems display 'anomalous' temperature distributions (Fig. 1) which are frequently ignored or imprecisely understood. Hence, hundreds of caves explored by speleologists to more than 500 m below the surface show temperatures close to those of the outside atmosphere (e.g. Choppy, 1984; Badino, 1992; Bitterli, 1996; Audra, 2001). As more than 10% of outcropping continental rocks are subject to intense karstic processes, interactions with human activities are frequent (Ford and Williams, 1989). A better understanding of heat fluxes in karst environments therefore seems necessary. This paper aims to provide new light on this problem.

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For the last 15 years, several authors have focused their research interest on a better understanding of heat flow processes within the saturated zone of karst systems (phreatic zone) (e.g. Benderitter *et al.*, 1993; Renner, 1996; Liedl *et al.*, 1997).

Owing to the high permeability of karst massifs, hydraulic gradients remain close to zero and water-tables are almost at the level of their outlets during low water episodes. The unsaturated (vadose) zone can therefore reach a thickness up to 2000 m. Only few authors have studied thermal processes in these thick vadose zones. Early attempts were made to predict cave temperature distribution theoretically (e.g. Eraso, 1965; Cigna, 1967; Andrieux, 1969; Wigley and Brown, 1971; Choppy, 1984) but most authors were only concerned with the first hundred metres of karst systems, where cave temperature is still influenced by seasonal variations. Recent publications have tended to neglect the role of the geothermal heat flux in the energy balance of such systems (e.g. Badino, 1995; Jeannin *et al.*, 1997; Lismonde, 2002), however the respective importance of heat fluxes transported by water and air is still a matter of debate. Field data relating to this issue are often missing. Observations presented in this paper aim at providing a new synthetic conceptual model of temperature distribution in a karst massif, based on available (largely unpublished) models and data.

Heat fluxes in a karst massif: a synthetic conceptual model

Basic assumptions

Karst systems under consideration are limited on their upper part by the outside atmosphere and on their lower part by low permeability (marls or unkarstified limestones). These aquifers can be subdivided into two major hydrological subsystems: the vadose (unsaturated) and the phreatic (saturated) zone (e.g. Ford and Williams, 1989). Both of these zones are subject to karstification processes, leading to the genesis of new conduits. Air and/or water circulations will therefore be observed in the active and fossil parts of a karst system (Fig. 2).

Heat transfer around the conduits, within the matrix, is considered to be restricted to conductive exchanges (i.e. water flow velocity in the matrix is low enough to be neglected) and tectonic complications leading to deep warm-water uplifts are not yet considered in this approach.

Thermal boundary conditions

Upper boundary: temperature at the surface

Air temperature at the surface of the karst massif is the result of a thermal balance between heat fluxes due to radiation, water precipitation, evapo-condensation processes and thermal exchanges between the soil and the air.

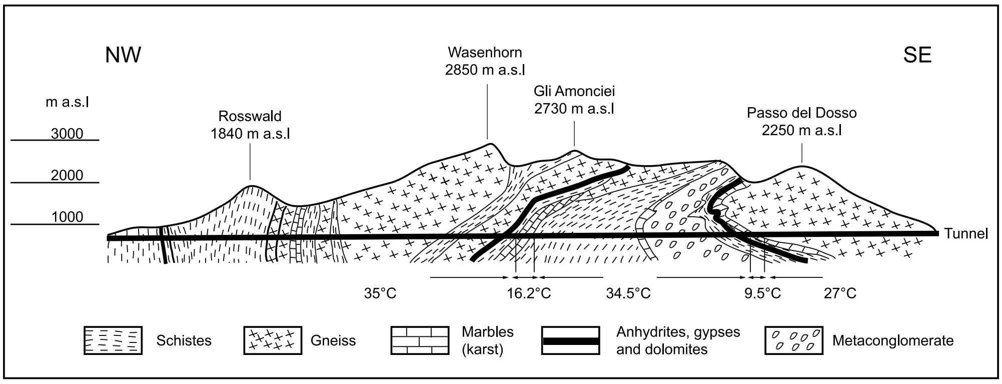

Fig. 1 Temperature profile in the Simplon tunnel (Switzerland). As in many other tunnels, thermal anomalies have been recognized in carbonate environments (figure adapted from Bianchetti *et al.*, 1993)

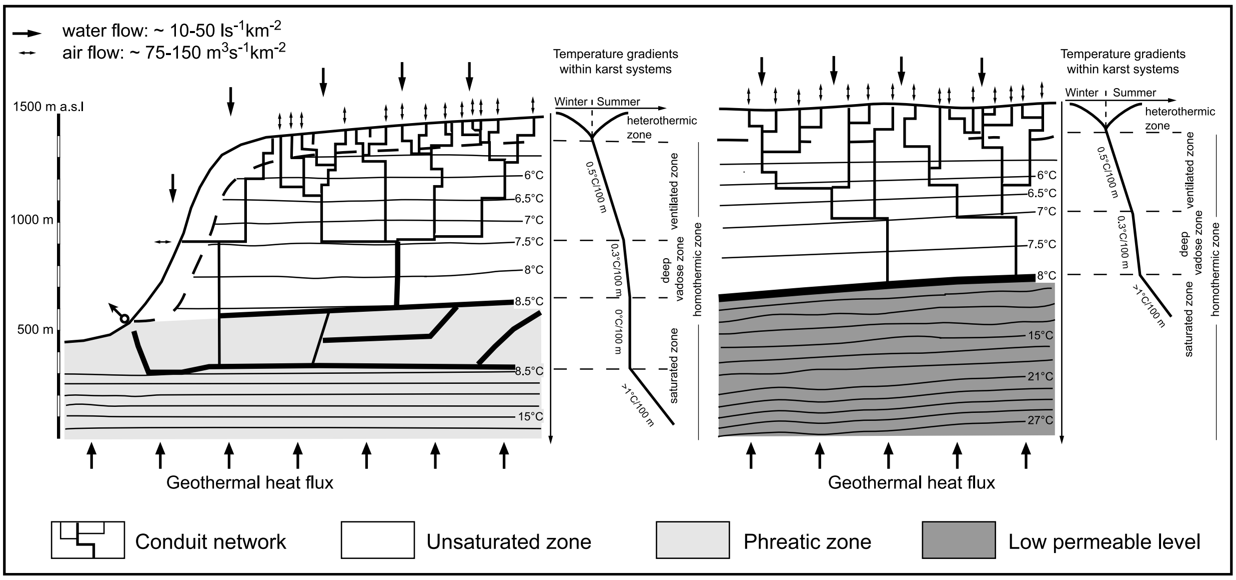

Fig. 2 Conceptual model of the temperature distribution in a karst aquifer. Close to the surface (i.e. for the first 50 m) seasonal variations are still observed. This is the heterothermic zone. Highly ventilated conduits located on the top of the unsaturated homothermic zone show steeper temperature gradients than the deep and poorly ventilated part of the vadose zone. Within the saturated zone, down to the bottom of the main conduit network, the gradient is close to zero. Below the main conduit system, temperature gradients are controlled by the geothermal heat flux.

In the atmosphere, air temperature varies linearly with altitude. Depending on air humidity, the observed gradient will vary between about -0.4 °C 100 m^{-1} (saturated air) and -1 °C 100 m^{-1} (dry air) (e.g. Triplet and Roche, 1986, p. 64).

Lower boundary: deep geothermal heat flux

In the continental lithosphere, the geothermal heat flux ranges between 40 and 140 mW m^{-2} depending on location on the Earth's surface (e.g. Hurtig *et al.*, 1991). As this flux largely results from radioactive decay within the Earth, steady-state conditions can be assumed.

It can easily be demonstrated that, owing to the specific discharge of karst springs in temperate regions (10–50 L s^{-1} km^{-2}), the energy resulting from the geothermal heat flux is mainly drained off and does not affect the temperature distribution in the unsaturated zone of a karst system (Mathey, 1974; Bögli, 1980; Drogue, 1985). The overall lower limit of the system can therefore be considered as a 'constant heat flow boundary', but this flux will hardly reach the unsaturated zone of the karst massif.

The heterothermic vadose zone

Thermal variations at the surface will affect the karst massif either by heat conduction through the matrix or by heat advection by water and/or air circulation. The upper part of the massif will therefore show annual temperature variations.

Based on numerous field observations (Table 1), the depth of this heterothermic zone is estimated at about 50 m, although some particular cases show values of more than 100 m. Unlike water and air flows, the conductive heat flux due to outside annual thermal oscillations cannot induce significant temperature variations at depths greater than 5 m. It can therefore be assumed that temperatures in the heterothermic zone are controlled by heat fluxes related to water and/or air circulation. This assertion is supported by the high porosity and permeability of the epikarst.

The homothermic vadose zone

The homothermic vadose zone is characterized by a high temperature stability: inversions of temperature gradients are not observed. Badino (1995) demonstrated the major role played by the rock heat capacity in this remarkable stability. Measurements have shown that rock, air and water are almost in thermal equilibrium although water and rock temperatures are always slightly lower than air (~0.15 °C, Jeannin, 1991). Observed gradients usually vary between 0.4 and 0.6 °C 100 m^{-1} (where z is positive with depth) (Table 1). An estimation is given below to assess the respective contribution of heat fluxes related to water and air circulation in the energy balance of the homothermic zone.

Heat fluxes due to water infiltration

Measured annual precipitation in temperate karst regions ranges between 500 and 2500 mm. Considering evapotranspiration, it is possible to estimate a specific infiltration of between 10–50 L s^{-1} km^{-2}, which is in good agreement with the discharge observed at most karst springs (e.g. Ford and Williams, 1989, p. 155).

The loss of potential energy of water during its vertical transit represents 9.81 J kg^{-1}m^{-1}. If this energy is fully transformed into heat by friction it leads to a temperature increase of 0.234 °C 100 m^{-1} (Lismonde, 2002). Considering the mean specific recharge (10–50 L s^{-1} km^{-2}), the heat supplied annually by the work of gravity to the homothermic zone of a 1000-m-thick karst massif ranges between 3×10^9 and 1.5×10^{10} kJ km^{-3}. Because thermal equilibrium in the homothermic zone is maintained, the heat advected by the water does not influence the vertical temperature distribution.

Heat fluxes due to air circulations

Considering the heat capacity and density of air and water, respectively, the volumetric air flow should be more than 4000 times higher than that of water in order to play a dominant role in heat exchanges. Mangin and Andrieux (1988) considered this unlikely and attributed the major heat transfer to be due to dripping water. The following section attempts to assess the order of magnitude of heat fluxes related to air circulation.

Air circulations in major karst systems are largely driven by pressure differences between the upper and lower entrances. In simple cases, this pressure difference can be approximated by (e.g. Trombe, 1952; Lismonde, 1981):

$$\Delta P_m \approx \frac{\rho_0 g h}{T_0}\left(T_i - \frac{T_A + T_B}{2}\right) \qquad (1)$$

where ΔP_m is the driving pressure (Pa), ρ_0 is the mean density of air (kg m^{-3}), g is the acceleration due to gravity (m s^{-2}), T_0 is 273 °K, T_i is the mean air temperature of the system (°K), T_A is the outside temperature at the top of the system (°K), T_B is the outside temperature at the bottom of the system (°K) and h is the altitude difference of the system (m); with an air flow given by the turbulent flow equation (Darcy–Weisbach):

$$q_m = \sqrt{\frac{|\Delta P_m|}{R}} \qquad (2)$$

where q_m is the air flow through the system (kg s^{-1}), ΔP_m is the driving pressure (Pa) and R is the aeraulic resistance of the conduit (kg^{-1} m^{-1}).

The aeraulic resistance (R) reflects the headlosses occurring in a natural conduit. Jeannin (2001) demonstrated that in karst conduits, regular headlosses dominate over singular headlosses. He provided measured values leading to an aeraulic resistance for conduits of 1 m diameter varying between 0.01 and 0.5 kg^{-1} m^{-1}, in agreement with estimates provided by Lismonde (2002).

Mean annual driving pressures of about ± 300 Pa are frequent in alpine karst systems. In such cases, a conduit with a simple geometry and a diameter of 1 m leads to an equivalent air flow of about 4.5 m^3 s^{-1}. Yet, our measurements confirm that air fluxes might be several times higher than that (e.g. Hölloch/CH $Q_{03.01.04}$: 15 m^3 s^{-1}; La Diau/F $Q_{20.02.04}$: 11 m^3 s^{-1}).

Worthington (1991) and Badino (1995) evaluated empirically the conduit density of a mature karst system at about 100 km km^{-3}, which is consistent with observations in known karst areas. Hence, an estimated air flow of 150 m^3 s^{-1} km^{-2} can probably be considered as a lower limit for a 1000-m-thick karst system. Given the temperature gradients of water and humid air (0.234 and 0.5 °C 100 m^{-1}, respectively), the heat flux associated with air circulation is assessed to be between 2 and 20 times larger than that of water, depending on recharge rates.

In many caves (Table 1), the observed temperature gradient is close to that of humid air (0.5 °C 100 m^{-1}). Therefore, it can be assumed that air circulation plays a dominant role in the temperature distribution in karst systems.

As observed in deep vadose zones (> ~500 m), or depending on local geological conditions, part of the unsaturated zone may be less connected to the land surface. Furthermore, the hierarchical structure of karst networks will concentrate into less numerous conduits at higher depth. As a consequence, air circulation due to forced convection will be progressively reduced. However, as water circulation is still present to the same magnitude, its relative effect on heat transfer might be significantly increased and temperature gradients will be much closer to 0.3 °C 100 m^{-1}.

Discussion

The suggested conceptual model is based on the assumption that the 'chimney effect' lies at the origin of most air circulation observed in alpine karst systems. It must be emphasized that further processes, such as barometric fluctuations, might lead to

Table 1 Temperature measurements in various karst massifs. Observations have shown that air, water and rock temperatures are always almost in equilibrium. Most observed gradients vary between 0.4 and 0.6 °C 100 m^{-1} (z is positive with depth; n.d., no data). (a) Measurements carried out in caves show the role of air circulation in the temperature distribution of the vadose zone; (b) data from boreholes show the influence of the geothermal heat flux below the main karst conduit network base level and of air circulation in the vadose zone

(a)

		Heterothermic zone	Ventilated system		Deep vadose zone		
Cave name	Climate	Depth(m)	Depth(m)	Air temperature gradient(°C 100 m^{-1})	Depth(m)	Air temperature gradient (°C 100 m^{-1})	Reference
Cathy (Switzerland)	Temperate	0–70	70–290	0.61	n.d.	n.d.	(this study)
Pleine Lune (Switzerland)	Temperate	0–180	180–230	0.59	–	–	(this study)
Bärenschacht (Switzerland)	Temperate	0–20	–	–	20–600	0.33	(this study)
Longirod (Switzerland)	Temperate	0–100	100–180	n.d.	180–450	0.38	(this study)
Mutsee (Switzerland)	Temperate	0–140	–	–	140–755	0.32	Weidmann (pers. comm)
Hölloch (Switzerland)	Temperate	n.d.	> 900	~0.5	–	–	Bögli (1980)
Rasse (France)	Temperate	0–100	100–470	0.43	n.d.	n.d.	Valton (pers. comm.)
Spluga della Preta (Italy)	Temperate	0–130	130–245	0.52	245–878	0.21	Bertolani (1975)
Tatras region (Poland)	Temperate	0–300	> 300	~0.5	–	–	Pulinowa and Pulina (1972)
Lechugilla (USA)	Semi-arid	–	0–300	0.55	–	–	(this study)
Kievskaya (Uzbekistan)	Semi-arid	0–100	100–300	0.78	300–420	0.33	(this study)
KT16 (Uzbekistan)	Semi-arid	0–25	25–150	0.1	150–200	0.18	(this study)
Gor Ulugh Beg (Uzbekistan)	Semi-arid	0–150	n.d.	n.d.	> 150	~0.35	Badino (1992)
Kijahe Xontjoa (Mexico)	Tropical	–	0–320	0.57	320–900	0.33	(this study)
Muruk (Papua New Guinea)	Equatorial	0–75	–	–	75–780	0.25	Audra (2001)

(b)

		Heterothermic zone	Ventilated system		Phreatic zone		Deep phreatic zone		
Cave name	Climate	Depth(m)	Depth(m)	Air temperature gradient(°C 100 m^{-1})	Depth(m)	Water temperature gradient(°C 100 m^{-1})	Depth(m)	Water temperature gradient (°C 100 m^{-1})	Reference
FM1 (Switzerland)	Temperate	0–60	60–440	0.61	n.d.	n.d.	n.d.	n.d.	(this study), Hessenauer *et al.* (2001)
FM2 (Switzerland)	Temperate	0–80	80–200	0.55	–	–	200–635	~1.5	(this study), Hessenauer *et al.* (2001)
Cachot (Switzerland)	Temperate	0–50	–	–	50–130	0.5–0.8	130–170	1.8	Mathey (1974)
Basse Fin (Switzerland)	Temperate	0–50	–	–	–	–	50–370	3.5	MFR Géologie-Géotechnique (2001)
St-Aubin 2 (Switzerland)	Temperate	n.d.	n.d.	n.d.	140–320	0.002	320–360	0.42	Bossy and Zwahlen (2000)

significant air flows in larger systems (e.g. Conn, 1966; Massen *et al.*, 1998). These will, however, increase the influence of air flow, as suggested by our model.

Nevertheless, several aspects are yet not included in this approach, which might in particular cases disprove the present conclusions. We summarize below the main controversial aspects and attempt to provide a theoretical explanation for them.

Geometrical factors

If the unsaturated zone is thinner than 100 m, air flow can be strongly reduced compared with the estimations provided by Eqs (1) and (2). Therefore, temperature gradients lower than 0.5 °C 100 m^{-1} could be observed. This situation is considered to occur in lowland plateau karst systems. However, this type of system usually receives little rain and the energy flux related to water is therefore also reduced. Temperature gradients close to 0.5 °C 100 m^{-1} are thus still possible.

If the system is thinner than ~50 m, it belongs to the heterothermic zone and gradients mainly depend on the outside temperature.

Climatic factors

If temperature variations at the surface are small (for instance in coastal or equatorial regions), temperature contrasts between cave air and outside air will be negligible. Thus, air flow due to forced convection will be reduced and temperature gradients will approach values dominated by water circulation (e.g. Muruk cave, New Guinea, Fig. 3).

Conversely, if water recharge is low, temperature gradients might be much closer to those of dry air (i.e. > 0.5 °C 100 m^{-1}). Unfortunately field data supporting this idea are lacking.

Hydrological factors

Owing to the reduced water circulation within confined aquifers, temperatures will be affected by the geothermal heat flux. Temperature gradients are therefore much higher than those observed in freely drained aquifers. This case is illustrated by temperature measurements taken from various boreholes (Fig. 4).

Conclusions

Heat fluxes due to air circulation in karst massifs are frequently underestimated. The physical concepts and observations presented here enable a better understanding of the temperature and heat flux distribution in karst environments. According to the proposed conceptual model, the following conclusions can be drawn:

1 air circulation defines temperature distribution in the homothermic unsaturated zone of karst aquifers in temperate climatic areas;
2 air temperature gradients approximately correspond to atmospheric gradient of humid air: 0.5 °C 100 m^{-1};
3 high water flows and/or low air circulation tends to reduce the

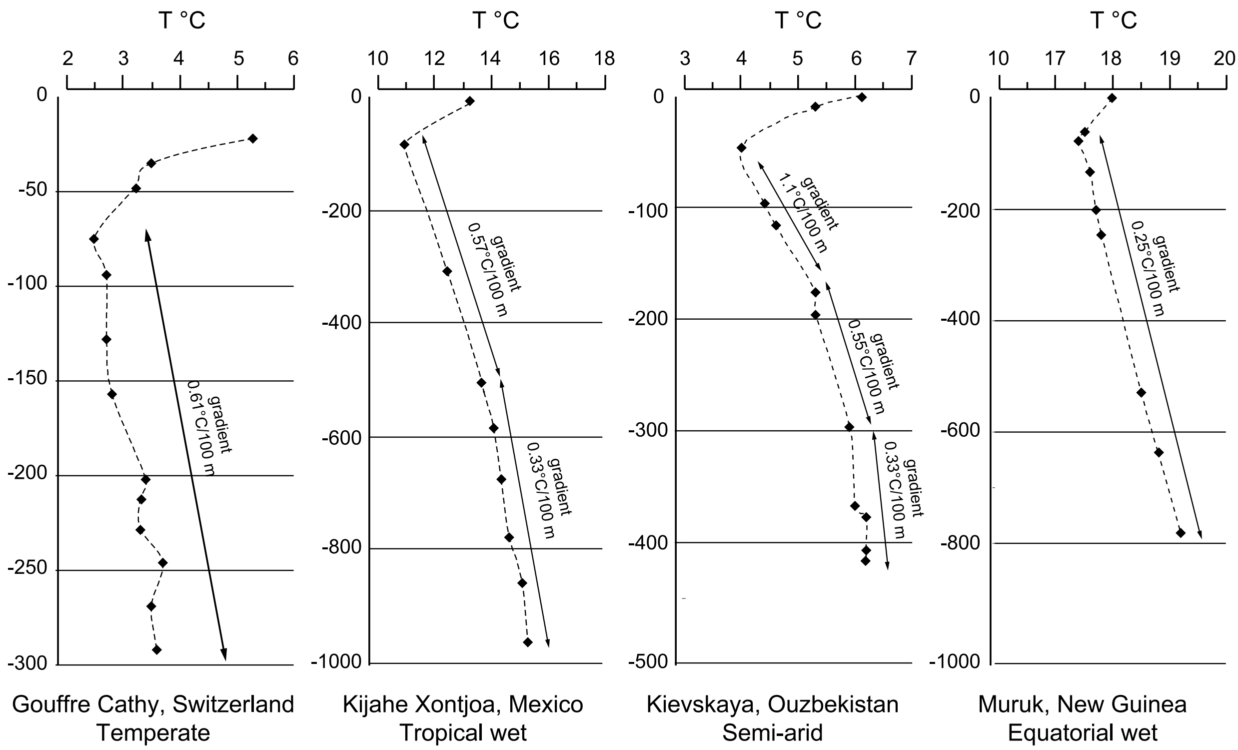

Fig. 3 Temperature measurements carried out in four caves under different climatic regimes (temperate, tropical, semi-arid and equatorial). Gradients in highly ventilated conduits are close to the humid air atmospheric gradient. In less ventilated zones, lower gradients are observed due to the relatively increased importance of water circulation. Data from Kievskaya show a temperature anomaly due to cold winter air circulation down to 160 m. Data from Muruk cave are taken from Audra (2001).

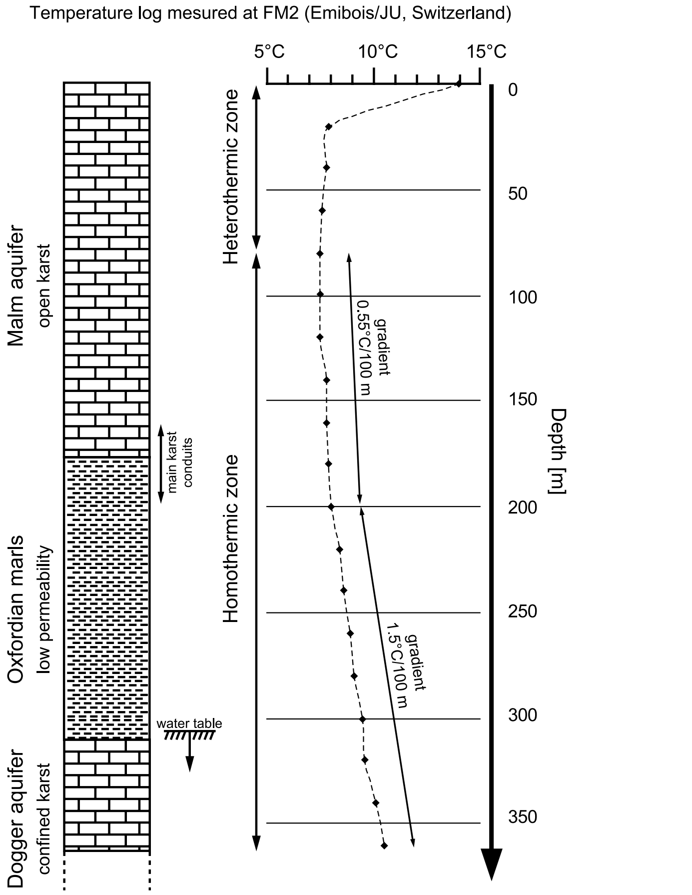

Fig. 4 Temperature measurements in a borehole of the Swiss Jura Mountains. Down to 175 m, the observed gradient (0.55 °C 100 m^{-1}) shows the influence of air circulation in boreholes, i.e. around the karst conduits. Below the main karst conduits (base of the Malm), a gradient of 1.4 °C 100 m^{-1} (instead of about 3.3 °C 100 m^{-1}) suggests that a slow water flow is present in the lower 'confined' aquifer of the Dogger limestone.

temperature gradient to about 0.3 °C 100 m^{-1};

4 water and rock temperature are close but slightly lower than air temperature;
5 temperature gradients between the main phreatic conduits and the top of the saturated zone are close to zero (Fig. 2);
6 temperature gradients below the main phreatic system are close to the normal geothermal gradient;
7 external climatic conditions define the significance of air and water heat fluxes.

The proposed conceptual model is built on numerous field observations. However, it has also been shown that heat fluxes might be significantly different depending on climate, geological context and maturity of a karst system. In order to validate the model, further temperature measurements in different karst regions are required. In addition, temperature logs in boreholes including measurements in the unsaturated zone will provide the necessary data for an extrapolation of these conclusions to all types of karst massifs.

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A process-based classification of alpine ice caves

Une classification des glacières alpines basée sur des critères climatologiques et glaciologiques

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Abstract

A short historical outlook provides a general overview of the processes at the origin of ice caves. A classification based on two criteria is proposed: 1) The cave air dynamics enables differentiation between cases where thermal trapping is at the origin of the ice (static ice cave) from those where a cold thermal anomaly is induced by a chimney effect (dynamic ice cave); 2) the type of ice enables one to distinguish endogenic ice caves (congelation ice) from exogenic ice caves (accumulation and transformation of snow). The intermediate types being frequent, the suggested classification consists of nine classes covering most ice caves of temperate regions.

Keywords : ice cave, definition, classification, cave climatology, cave glaciology.

Résumé

Un aperçu historique et un survol des processus à l'origine des glacières permet de proposer une classification selon deux critères: 1) La dynamique de l'air de la grotte distingue les cas où un piège thermique (glacière statique) est à l'origine de la glacière des cas où un effet cheminée créé une anomalie thermique froide (glacière dynamique); 2) Le type de glace permet de distinguer les glacières endogéniques (glace de congélation) des glacières exogéniques (accumulation et transformation de neige). Les cas intermédiaires étant fréquents, la classification proposée contient neuf classes couvrant la majorité des glacières situées en régions tempérées.

Mots clés: Glacière, définition, classification, climatologie souterraine, glaciologie souterraine.

1 | Introduction

Since subsurface ice occurences (i.e. in caves) were first mentioned, a specific vocabulary has emerged in the common language. This highly descriptive terminology (e.g. Creux de glace, Schneeloch, freezing cavern, etc.) has a regional value, but the multiplicity of terms leads to a general misunderstanding in the scientific language. It is not the aim of this paper to define or influence the regional vocabulary, but rather to define a scientific nomenclature which can be widely used and understood.

Actually, numerous scientists investigated caves with ice and introduced new definitions according to their respective sites and observations. Thury (1861) first suggested a classification based on his understanding of the processes leading to the presence of cave ice. Later, morphological studies of cave ice led to the distinction of several ice patterns (e.g. Kyrle, 1923).

However, accessibility to the early publications was limited and investigations of the respective authors remained mostly ignored. Therefore, noteworthy differences occurred according to the various languages. A first analysis of numerous articles published over the last several decades outlines a heterogeneous (if not contradictory) nomenclature (Luetscher & Jeannin, 2001).

Due to the increasing interest in karst records, new studies were initiated since 1990. Most of the studies attempt to better understand the processes at the origin of subsurface ice fillings and/or to investigate the potential paleoclimatic archives contained in massive cave ice deposits, consisting in larger perennial firn accumulations or congelation ice (antonym: pore ice). In this context, a unification of the terminology related to subsurface ice occurrences is necessary for the presentation of results and ideas about ice caves.

This paper aims at proposing a definition of «ice caves». It is organized in two parts. First, a short summary of the known nomenclature is presented along with observed contradictions, explained and argued in the light of recent advances in this field of research. Then, a process-based classification of alpine ice caves is sketched, as close as possible to the existing terminologies but reorganized in a consistent framework.

2 | Outlook on the existing nomenclature

2.1 Definition of «ice cave»

Balch (1900) first outlined that in English, the term «ice cave» is somewhat misleading since the character of the content is mentioned before the nature of the geological formation. This author also emphasized that the term implicitly excludes mines, tunnels, boulder talus, etc. Unfortunately, Balch's proposition to use the French term «Glacière» had only little success and most of the recent English textbooks (e.g. Gunn, 2003; Ford & Williams, 1989; Bögli 1980) still use the term «ice cave». Therefore, in accordance with the 1st International Workshop on Ice Caves (Capus, Romania, 2004) and with the most popular textbooks, the following definition is proposed:

«Ice caves» are rock-hosted caves containing perennial ice or snow, or both.

Ford and Williams (1989) also include seasonal ice fillings in this definition. Although this makes sense from a climatological perspective, the most common opinion is to consider only perennial ice fillings. Bögli (1980) would prefer to consider snow fillings separately. Yet, it seems justified to maintain the largest possible definition and to define more specific terms in the classification of different types of ice caves.
According to the Cryokarst Commission of the International Union of Speleology (IUS), voids resulting from channelled water circulations below glaciers are referred to as «glacier caves» (e.g. Smart, 2003).

2.2 Types of ice caves: some theoretical aspects

Cave ice has been observed and thoroughly described in numerous caves around the world (e.g. Hill & Forti, 1997). It is not the aim here to provide an extensive review of ice types found in caves, however, it should be noted that at least seven different types have been recognized (Ford & Williams, 1989). Table 1, adapted from Ford & Williams (1989), Hill & Forti (1997) and Yonge (2003), lists most of these cave ice patterns and refers to the major studies dedicated to them. According to the existing literature and based on our own observations cave ice mainly results from recrystallisation of snow, freezing of infiltration water (ponded ice, drip or flowstones) or deposition of hoar frost. Furthermore, intrusive ice can be observed in the proximity of temperate glaciers (e.g. Castleguard, Ford et al., 1976) and extrusive ice flowers have been described in several caves (e.g. Hill & Forti, 1997).
Cave ice can be found either in permafrost or temperate regions, usually in the vicinity of cave entrances where temperatures are influenced by seasonal variations (heterothermic zone). As only a few caves are known in the continuous permafrost zone, the present paper focuses on temperate regions wherein cave ice could be described from all around the world (e.g. Maire, 1990). Thury (1861) was one of the first authors to outline the processes at the origin of low altitude cave ice in temperate regions, which were later concisely described by Balch (1900). In these regions, temperature anomalies leading to the freezing of water either result from thermal trapping due to one single entrance or from forced convection (chimney effect) related to multiple entrances. Hence, a classification of ice caves based on cave air dynamics (CAD) criteria could be proposed. At one end of the criteria, static ice caves are related to «cold air trap» situations and at the other end, dynamic ice caves are related to «chimney effects».
According to Thury (1861), static ice caves are those where summer air circulations are mostly insignificant. This situation occurs in downward sloping conduits that are closed at the bottom. Because of density differences between cave air and outside air, major air circulation occurs during an «open period» (Girardot & Trouillet, 1885) corresponding mostly to the winter season, when $T°_{outside} < T°_{cave}$. Although it is easy to demonstrate that natural convection cells still subsist during the summer season, they are not relevant for energy exchange with the outside atmosphere. Therefore, cold temperatures are maintained within the cave.
In dynamic caves, according to Thury (1861), chimney effect is the predominant process at the origin of the cooling. It occurs in caves with two or more entrances located at different elevations; due to pressure differences, and induced by the density difference between cave air and outside air, significant unidirectional air circulations take place between the various entrances. Since air temperature in caves is nearly constant (e.g. Wigley & Brown, 1976; Choppy, 1984; Lismonde, 2002), while external temperatures vary, chimney-effect winds reverse direction seasonally. During the winter season cold air is sucked in at the lower entrance and enables the freezing of infiltration water there. This results in a cold thermal anomaly at the lower entrance. Conversely, a «warm» anomaly can be observed at the upper entrance.
The term «statodynamic ice cave» was introduced later (for instance, Bögli, 1980) for caves of an intermediate type falling between the two categories described by Thury (1861). However, several authors outlined that this classification based on CAD does not provide any indication about the origin of the ice. Among others, Maire (e.g. 1980; 1990) distinguished ice caves based on the genesis and dynamics of their ice fillings. He described three major types of cave ice: 1) intrusive glacier ice; 2) congelation ice; and 3) firn. Unfortunately, this classification also uses the terms static-, statodynamic- and dynamic- ice cave, initially introduced for a climatic characterisation of ice caves.

Exogenous cave ice :

Ice type	Description	Formation	Litterature
Firn (recrystallized snow)	Opaque to bluish, layered.	Accumulation of snow in cave traps which densifies and recrystallizes with infiltration component.	e.g. Maire (1990); Bini & Pellegrini (1998).
Intrusive	Massive ice subliming on the caveward face or with hoarfrost formation.	Glacier ice intruded into cave passages at the glacier/rock contact.	e.g. Ford et al. (1976).

Endogenous cave ice :

Ice type	Description	Formation	Litterature
Drip or flowstone (congelation ice)	Stalactites, stalagmites, flowstone. Polycristalline, clear to opaque.	Freezing of infiltration water.	e.g. Racovita (1994); Pulinowa & Pulina (1972); Kyrle (1923; 1929); Viehmann & Racovita (1968).
Ponded ice (congelation ice)	Clear or coarse polycrystalline ice. Occasionally with bubbles.	Static water that freezes from the top downward. Can incorporate infiltration water and falling hoar ice.	e.g. Marshall & Brown (1974).
Hoar frost	Needles, rosettes and hexagonal plate crystals up to 0.5 m. Small, tapering prismatic crystals.	Humid air condensing onto cave walls below 0 °C.	e.g. Lauriol & Clark (1993); Waldner (1933); Halliday (1966).
Ice in clastic sediments	Intra-particle aggregations as irregular masses, lenses, needles or soil-like extrusions.	Freezing of moist sediments.	e.g. Pulinowa & Pulina (1972); Schroeder (1977).
Extrusive	Curving fibrous crystals, similar to gypsum flowers up to 20 cm long.	Supercooled water forced through microfissures in rock below 0 °C which freezes as it emerges from the cave walls.	e.g. Ford & Williams (1989).

Tab. 1 Perennial ice occurrences in caves (adapted from Ford & Williams, 1989; Hill & Forti, 1997; Yonge, 2003).

Tab. 1 *Type de glace trouvée dans les grottes (d’après Ford & Williams, 1989; Hill & Forti 1997; Yonge 2003).*

3 | The suggested classification

The classification suggested in this article is restricted to caves with massive ice occurrences, that is firn and congelation ice fillings (Shumskii, 1964), located in temperate regions (i.e. «alpine» type). Caves with perennial hoar frost, frozen sediments and ice extrusions are not considered here. Furthermore, because of the very restricted number of known examples, intrusive glacier ice is considered as a separate category.

Two major classification systems are currently used for the characterisation of ice caves. The classification based on CAD processes, introduced at the end of the nineteenth century, provides interesting elements for paleoclimatological investigations in cave ice. Therefore, it seems meaningful to keep these aspects in the proposed nomenclature. Nevertheless, a distinction between different types of cave ice provides morphogenetic information. Distinguishing between «endogenic» ice patterns (e.g. congelation ice) and «exogenic» ice (i.e. snow/firn accumulations) also provides useful information for thermodynamic considerations.

Figure 1 suggests a synthesis of both classifications for the distinction of alpine-type ice caves. This classification is primarily based on CAD criteria and is consistent with the terminology introduced by Thury (1861), which is still in use in most of the English literature. Hence, the terms static, statodynamic and dynamic, refer to the presence or absence of forced convection (static for air traps and dynamic for chimney effects). A second criteria which includes the ice formation processes (firn accumulation versus congelation ice) is also considered, incorporating concepts from the French literature (Maire, 1980; 1990).

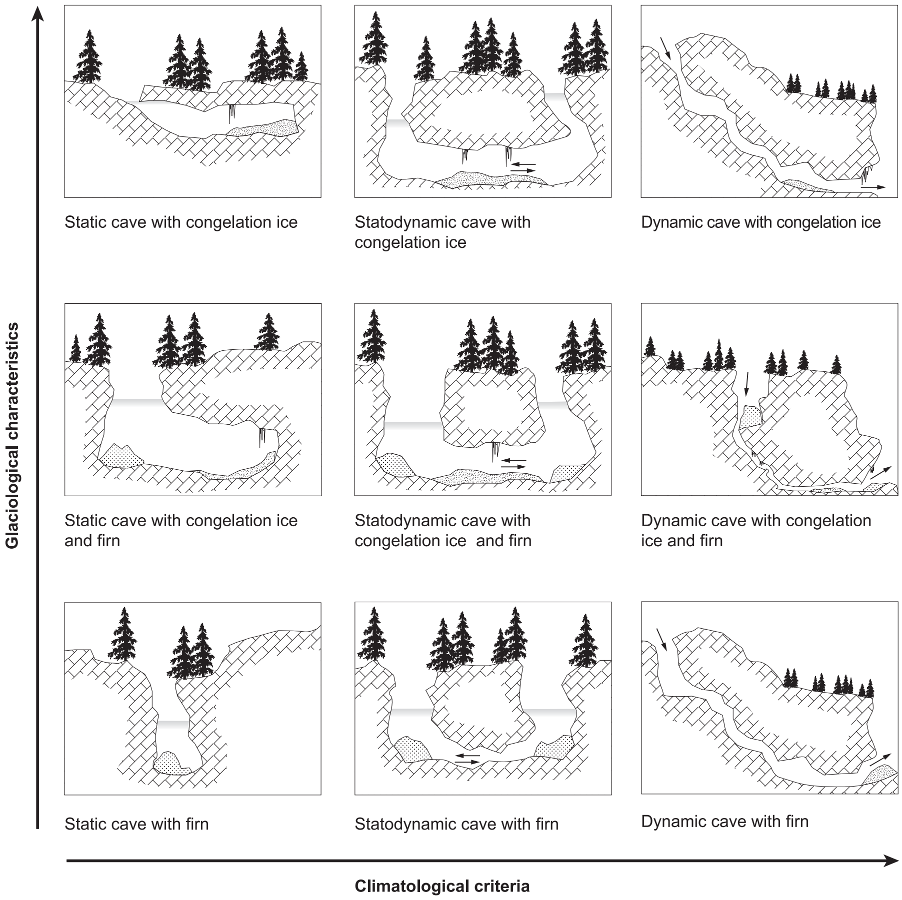

Fig. 1 Suggested classification of ice caves. Two criteria are considered in this classification: 1) On the horizontal axis, cave air dynamics distinguishes caves where thermal trapping is at the origin of ice from those where a cold thermal anomaly is induced by chimney effect; 2) On the vertical axis, ice types are distinguished between exogenic ice (i.e. firn accumulation) and endogenic ice (i.e. congelation ice).

Fig. 1 *Proposition de classification pour les glacières. Deux critères sont considérés: 1) Sur l'axe horizontal, la dynamique de l'air souterrain distingue les grottes où la glace se forme grâce à un piège thermique de celles où l'anomalie thermique froide résulte d'un effet cheminée; 2) sur l'axe vertical, on distingue la glace exogénique (i.e. accumulation de névé) de la glace endogénique (i.e. glace de congélation).*

4 | Discussion

The authors consider that a classification of ice caves based on cave air dynamics (CAD) is meaningful for two main reasons:

1. it is close to the most widely used classification;
2. it reflects the dominating process at the origin of the ice, which is a prerequisite for any further investigation of such a cave.

Nevertheless, glaciological criteria must also be considered as it enables one:

1. to determine the origin of ice and thus to better anticipate glaciochemical responses related to the type of ice;
2. to assess the contribution to the energy balance of the ice cave.

Simplifying the origin of massive ice into two classes (i.e. firn or congelation ice) could introduce some confusion. It must be noted that this classification describes the origin of ice and not that of the water, which can vary from cave to cave (i.e. infiltration water, phreatic water, melted snow, etc.). Determining the latter may be difficult to establish in the field. Glaciochemical signatures and crystallographic observations should enable distinction between different origins of water in congelation ice, but contamination due to remobilisation after melting is still possible. Therefore, a more precise classification, although better from a theoretical point of view, would be inapplicable in reality.

Processes at the origin of the proposed classification are focused on ice caves located under temperate climates. In continuous permafrost regions, below the active layer (superficial layer above permafrost which thaws during summer), ice in caves is rather a rare phenomenon due to the absence of water (e.g. Ciry, 1962). However, seasonal air circulations are possible leading to the freezing of hoar frost. Conversely to temperate regions, the deposition of hoar frost is not related to the presence of cold temperature anomalies but rather to the possible presence of «warm» and humid air circulations. This positive anomaly can occur in cases of a chimney-effect wind circulation in the upper entrance zone or in an upward sloping conduit acting as a thermal trap.

Regarding paleoclimatic studies, the CAD-based classification supports the consideration of CAD in the scientific studies. A good understanding of ice cave processes enables one to link the presence of cave ice to outside climatic conditions, especially to some specific parameters (e.g. freezing index). Also, the spatial distribution of ice caves can be better understood.

Moreover, seasonal ice fillings are not considered yet in this classification in accordance with the opinion of several ice-cave scientists. Nevertheless, it is easy to demonstrate that the processes at the origin of seasonal ice patterns remain the same as those at the origin of perennial cave ice. The only difference is that the overall energy balance is positive, leading to the melting of ice. This situation results either from the implicit morphological characteristics of the cave or its surroundings but can also be induced by the effect of regional climate. Considering climate changes during the Holocene, it can be estimated that the minimal elevation of ice caves rose several hundred metres during this period. This means that seasonal ice caves and even caves without ice may have been real ice caves in the past. Signatures of paleo-ice fillings might have been preserved in cave sediments, making it possible to reconstruct the past distribution of ice caves, which is an interesting potential for paleoclimatic reconstructions. Therefore, even former or seasonal ice caves can be interesting and should be documented. The proposed classification could be expanded further in order to include this temporal aspect of ice caves.

Although this classification has demonstrated its limits, it is believed to be a first step towards a terminology of ice caves common to all scientists involved in this field of research. Further discussions will certainly help to improve this proposition.

5 | Acknowledgements

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Ice caves as an indicator of winter climate evolution

– a case study from the Jura Mountains – published in: *The Holocene,* **15**(**7**): 982-993

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Abstract

Subsurface ice fillings were first described in the Jura Mountains at the end of the sixteenth century. In order to assess the impact of climate change on low altitude cave ice a detailed inventory has been drawn up and more than 50 objects have yet been identified. Comparisons between older cave maps, photographic documents and present day observations, outline a negative trend of the ice mass balances, which increased at the end of the nineteen eighties. As most of these ice caves act as cold air traps, this negative mass balance is mainly attributed to higher winter temperatures and to reduced snow precipitation at low altitudes. The equilibrium line altitude of ice caves is supposed to have increased several hundred meters between 1978 and 2004. Photographic comparisons and proxy records in some of the studied caves provide evidence of a quick mass turnover. Ice ages range between less than a few decades and a Millenium. Climatic records in these ice fillings will therefore present only short time series compared to other cave sediments. However, indications of former ice fillings have been found in different caves of the Jura Mountains and outline their potential role as paleoclimatic markers.

Key words: ice cave, climate, Jura mountains, mass balance, freezing index.

1 | Introduction

During the twentieth century, climate changes induced a negative trend of the mass balance of most glaciers in the alpine belt (Hoelzle et al., 2003; Braithwaite, 2002). Summer air temperatures and precipitation regimes (taking implicitly into account net radiation and albedo variations) have been recognized as being among the most important factors influencing glacier mass balances (e.g. Paterson, 1994). Therefore, glacier fluctuations are commonly considered to represent key factors for the early detection of enhanced greenhouse effects on climate (Kuhn, 1980; Haeberli et al., 1999). The mass balance of subsurface ice fillings, which can be found in low-altitude regions and more particularly in rock-hosted caves, is by far less known. Recognized for a long time for their natural interest, such features have been defined in English literature as «ice caves» (e.g. Ford and Williams, 1989), which must not be confused with «glacier caves», resulting from channelled water circulations in glaciers (e.g. Smart, 2004). The present paper focuses on the way ice masses in caves evolved during the last few centuries and attempts to link their evolution to outside climatic parameters. If this link can be established, ice caves could be used for paleoclimatic reconstructions.

Ice caves have been described in many parts of the World at altitudes where the mean annual air temperature is several degrees above 0°C (e.g. Harris, 1982; Maire, 1990). Early investigations (e.g. Thury, 1861; Balch, 1900) provided qualitative descriptions of the processes at the origin of these sporadic permafrost occurrences (Haeberli, 1978). Several authors pointed out the importance of cave air circulation for the cooling of such systems in temperate regions (e.g. Bock, 1913; Luetscher and Jeannin, submitted a). In winter, due to the density difference between «warm» cave air and «cold» outside air, significant air circulation (up to 20 m^3/s) takes place throughout the cave. This air flow enables substantial heat exchanges leading to the freezing of any available infiltration water as well as to the conservation of intrusive snow accumulation and to evapo-condensation processes (i.e. sublimation and/or hoar frost deposits, for further details see also Yonge, 2004; Hill and Forti, 1997). Thus, a new layer of ice is formed each winter at the top of the ice volume leading to its stratification. Conversely, heat exchange with the outside atmosphere is considerably reduced during the summer season when cold air trapping or reversed ventilation occurs (e.g. Luetscher and Jeannin, in press, a). Due to this limited heat exchange and to the high latent heat of ice, long-term preservation of subsurface ice fillings is possible. The high efficiency in the energy exchange during the winter season suggests that ice mass balance is closely related to outside winter climatic conditions.

Despite the well known climate change (IPCC, 2001), only few authors provided quantitative data about recent cave ice fluctuations. Based on a detailed literature review and field observations in the Jura Mountains, the present study aims at giving facts and indications about fluctuation of the cave ice mass balance and their relation to the evolution of the outside climate. Although this paper is mainly based on preliminary results obtained from the compilation of scattered historical data, interesting data could also be obtained from a three-year observation carried out in two selected ice caves.

Such data constitute the indispensable background for the assessment of the potential paleoclimatic records in ice caves. Hence, ice caves could provide data in regions where other records are missing (e.g. Perroux, 2001). However, first observations indicate a fast mass turnover and expected records will cover a period ranging between 100 and 1000 years. For older records, the study of cave sediments can provide precious information on past winter climate conditions.

2 | Site description

The Jura Mountains form a ca. 400 km wide arc between the Savoie (France) in the South-West and the Black Forest (Germany) in the North-East (fig. 1). The inner part of this NW arcuate range, mostly located in Switzerland, is characterized by a succession of crests and valleys ranging between 1000 and 1500 m a.s.l., with the highest peaks reaching about 1700 m a.s.l. The external Jura is characterized by flat plateaus, limited to the North and separated from each other by numerous small scale tear faults. The mean elevation in this area is about 700 m a.s.l. The Jura Mountains consist essentially of sedimentary rocks, presenting a stratigraphy of alternating limestones and marls up to 1000 m in thickness. Dominated by carbonate rocks, most of the Jura Arc is affected by an intense karstification process. The temperate climate of the Jura Mountains is strongly influenced by oceanic meteorological conditions. As a first obstacle from the atlantic ocean, the Jura Mountains benefit from abundant precipitations, usually between 1200 and 1600 mm/year, which can reach more than 2000 mm/year on certain crests.

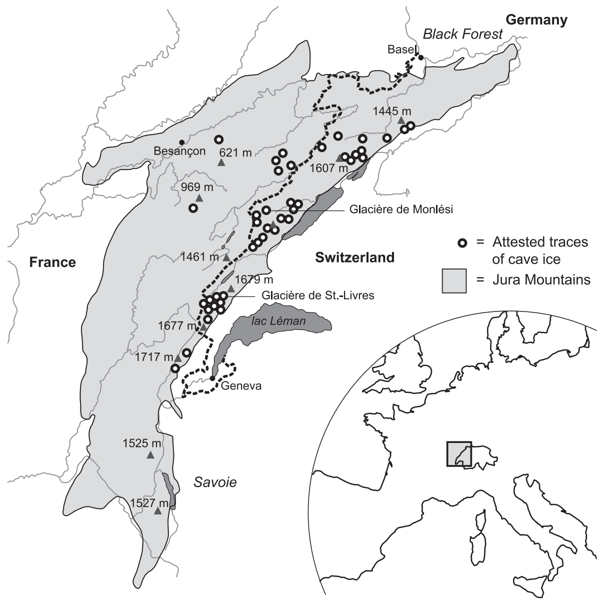

Fig. 1 Geographical situation of the study site. Of over 10,000 caves explored and documented in the Jura Mountains only a few dozens contain a perennial ice filling. These ice caves are mostly located in the inner part of the Jura range, where the highest peaks and the harshest winter conditions prevail.

In general, the western side presents wetter conditions than the eastern one, which is already characterized by a semi-continental regime. However, the yearly precipitation distribution over the entire massif seems to be quite random and long-term measurements show a pluviometry

varying locally between oceanic and continental regimes (Gaiffe, 2001). Snow precipitation occurs about 50 days/year on the higher summits and represents half of the total precipitation observed. The accumulation can reach more than 2 m and remains from November to May in certain combes of the high range.
Because of karstification, groundwater recharge represents between 50 and 75% of the total pluviometry, depending on the elevation (e.g. Tripet, 1973; Jeannin and Grasso, 1995; Luetscher and Perrin, accepted). Most of this infiltration takes place in late autumn and in April during snowmelt. In summer, major precipitation events related to strong thunderstorm can easily represent about 20 litres/m^2 during a short period (less than one hour). Temperatures present large contrasts between the warmer and the colder months. For instance, the maximal annual amplitude reaches up to 50 °C while a difference of 17 °C between the warmest and the coldest months is common. According to air temperature records (Meteoswiss), an average vertical gradient of -0.5 °C/10 m can be estimated which is close to the adiabatic lapse rate. The altitude of 0 °C-isotherm can therefore be extrapolated to approximately 2200 m a.s.l.
Due to its topography, many cold air basins can be observed during the winter season, creating frequent thermal inversions. It is therefore quite frequent to observe warmer temperatures on the slopes than in the valleys.

3 | Methods

During the last hundred years, speleologists explored and documented over 10,000 caves in the Jura Mountains (e.g. Audétat et al., 2002; Bitterli, 1996; Chirol, 1985; Gigon, 1976; Gigon and Wenger, 1986; GIPEK, 1988, 1991, 1996, 2004). Exploration reports are usually provided in the form of descriptions and topographic cave maps. Literature and speleological archives have been compiled in order to identify caves where perennial or temporary ice/firn accumulation might occur.
Data consist mostly of qualitative observations while in some specific cases ice volumes were estimated based on morphological criteria. Their compilation had to face problems related to similar names for different caves (for instance, Creux-de-Glace), different names for the same cave (for instance, Glacière Pierrette / Baume ouest du Petit-Pré) or typing errors in the mentioned coordinates. Furthermore, several objects were described as ice caves but probably never contained more than a short seasonal snow accumulation (for instance, Glacière des Raisse-Gueissaz).
Field verifications were carried out in order to provide a validated list of caves in the Jura Mountains which might (have) contain(ed) a perennial ice filling. Based on morphological criterion, selected caves have been visited and described in detail with the purpose of acquiring a good overview of the 2001-2004 situation. Maps of visible ice fillings have been sketched using traditional speleological mapping methods (e.g. Grossenbacher, 1991). Observations of air circulations, water infiltrations and temperature measurements have been systematically carried out and ice fillings have been distinguished between congelation ice and firn accumulation. Although major uncertainties remain about the thickness of the ice filling, ice volumes were estimated based on the observed morphologies. Despite an inaccuracy of nearly 30% on the ice volume assessment, simple tape readings provided good estimations of maximal ice extents. Results were then compared to historical data, but also provided a precious reference point for future comparisons.
To facilitate the understanding of long term evolutions, detailed field studies were carried out in two of the largest ice caves (Glacière de Monlési and Glacière de St-Livres). Both of these sites were thoroughly documented and described. Datings of proxy records were attempted in order to acquire further information on mass turnover rates. Furthermore, a three year topometric survey, using reference points marked in the ice filling, was carried out. The accuracy of measurements is in a range of ± 1 cm which is more than enough for the order of magnitude required for the assessment of ice fluctuations and mass turnover. Meteorological data used for correlations with outside climate were provided by Meteoswiss. Precipitation and air temperature records are carried out, among others, at several stations in the Jura Mountains. Monthly air temperature time-series were reconstructed and homogenised up to 1865 for the station of Chaumont (6°59'16.3''E / 47°3'4.7''N / 1073 m a.s.l.). Snow precipitations and daily temperature values, available from this station since 1959, were compared for consistency and representativeness with other stations in the Jura Mountains.

4 | Results

4.1 Characteristics of ice caves in the Jura Mountains

Caves of the Jura Mountains are mainly located in jurassic limestones and belong to larger karst systems which are subject to major water and air circulations. In these systems, the strong inertia provided by the surrounding rock is responsible for a very stable cave air temperature. Also, temperature distribution in the Jurassic vadose karst systems is close to the natural gradient of humid air (Jeannin et al. 1997; Luetscher and Jeannin, 2004) and values are usually comparable to the yearly outside average temperature at the same elevation. The zero degree isotherm being located far higher than the summits of the Jura Mountains, freezing temperatures are not expected in caves. However, special conditions can lead to the presence of cave ice. Note that freezing conditions are restricted to local parts of the caves (entrance), which are still influenced by seasonal variations (e.g. Choppy, 1984).

Speleological data compilation and field verifications identified 25 caves presenting perennial ice (s.l.) occurrences (tab. 1a). Mostly located in the inner range of the Jura Mountains, they are all at altitudes of above 1000 m and always consist of downward sloping caves generally leading to an obstructed conduit. For caves with more than one entrance, altitude difference between them is usually within a few meters. Therefore, most ice caves in the Jura Mountains act as cold air traps (e.g. Luetscher and Jeannin, submitted b). The only noticeable exception is the Glacière de Druchaux, situated at 1495 m a.s.l., which shows a chimney effect between the main shaft and a doline located a few meters higher than the cave entrance.

Large entrance shafts (diameter of > 5 m) enable intrusive snow accumulations in many caves but congelation ice can also be observed. Observed ice volumes vary from a few cubic meters to more than 6,000 m^3 in Monlési ice cave (Luetscher and Wenger, 2002).

4.2 Dynamics of cave ice

4.2.1 Yearly ice accumulation

Perennial ice patterns in caves of the Jura Mountains result from the diagenesis of snow accumulated during the winter season and/or from the freezing of infiltration

Tab. 1a Selected ice caves in the Jura Mountains. 25 perennial ice fillings could be identified in caves (table 1a). Perennial ice consists mostly of firn accumulations and sometimes of congelation ice. Estimated ice volumes vary between a few cubic meters to over 6000 m^3 in Monlési ice cave. Evidence of former ice filling could be found in numerous other caves (table 1b). (n.d.: no available data)

	Name	E	N	Z	Number of entrances	Ice Depth	Cave morphology
1	Glacière de Monlési	6°35’3.5’’	46°56’17.6’’	1135	3	-20 to -33 m	room
2	Glacière de St-Livres	6°17’46.3’’	46°33’55.0’’	1362	1	-16 to -45 m	desc. conduit
3	Creux de Glace de Courtelary	7°4’52.7’’	47°9’10.5’’	1330	1	-25 to -32 m	room
4	Glacière du Crêt des Danses	6°8’6.6’’	46°30’12.6’’	1490	1	-28 to -34 m	shaft
5	Gouffre de Bellevue	6°40’43.2’’	46°54’8.4’’	1348	1	-21 to -31 m	shaft
6	Creux-Bastian	6°41’26.8	46°55’4.7’’	1210	2	-25 to -27 m	shaft
7	Baume à la Neige	6°18’30.4’’	46°35’18.7’’	1560	1	-17 to -34 m	room
8	Glacière sud (2) du Mont-Tendre	6°18’59.8’’	46°54’8.4’’	1575	1	-12 to -15 m	shaft
9	Gouffre du Mont des Verrières	6°28’54.1’’	46°53’9.8’’	1189	1	-40 m	shaft
10	Gouffre 1 des Grands Bois	6°28’15.8’’	46°52’56.0’’	1153	1	-20 to -31 m	shaft
11	Glacière de la Pierre-à-Coutiau	6°17’5.7’’	46°34’51.9’’	1575	1	-16 to -23 m	shaft
12	Creux à la Neige	6°8’47.5’’	46°35’37.8’’	1410	1	-10 to -15 m	shaft
13	Baume de la Passoire	6°18’37.1’’	46°35’22.2’’	1550	4	-10 to -30 m	shaft
14	Glacière du Couchant (1)	6°9’48.6’’	46°31’4.1’’	1470	1	-29 to -31 m	shaft
15	Baume du Bois des Begnines (2)	6°9’40.3’’	46°30’55.4’’	1510	1	-13 to -15 m	shaft
16	Glacière du Couchant (2)	6°9’37.4’’	46°30’21.2’’	1430	3	-10 to -14 m	shaft
17	Glacière de Druchaux	6°18’16.8’’	46°34’53.8’’	1495	1	-30 to -100 m	shaft
18	Glacière de St-George	6°14’25.9’’	46°31’33.0’’	1290	2	-20 to -25 m	room
19	Gouffre des Croix-Rouge (1)	6°9’34.8’’	46°30’51.6’’	1520	2	-30 to -35 m	shaft
20	Glacière Paul Matile	6°50’20.6’’	47°8’31.2’’	980	2	-1 to -3 m	vert. fracture
21	Glacière des Amburnex	6°13’24.5’’	46°33’13.8’’	1280	2	-4 m	conduit
22	Puits à Neige du Mont Sallaz	6°9’12.5’’	46°30’1.2’’	1430	1	-5 to -7 m	shaft
23	Gouffre Nord des Cailles	5°57’51.3’’	46°17’25.8’’	1380	1	-44 to -49 m	shaft
24	Glacière du Col du Crozet (sup)	5°57’51.3’’	46°17’25.8’’	1545	1	-12 to -25 m	shaft
25	Neigière d’Arc-sous-Cicon	6°24’8.9’’	47°1’45.4’’	1070	1	-20 m	doline

water (congelation ice, Shumskii, 1964). Both processes preferentially take place during early spring when thawing of exterior snow deposits begins. Temporary ice patterns like stalagmites and stalactites grow below some chimneys or water inlets and were the object of thorough descriptions in different ice caves around the world (e.g. Racovita and Onac, 2002; Kyrle, 1923). Although melting quickly affects such small features, part of the ice deposited during winter can be preserved the year-round. Estimation of seasonal ice accumulation in Monlési ice cave provided values ranging between 100 and 500 kg/m^2, i.e. ~ 10 to 50 cm of ice per year (Luetscher et al., 2003).
In most cold air trap caves, due to the lack of significant air exchanges with the outside atmosphere during the summer season, seasonal melting is generally related to water infiltration (e.g. thunderstorms). Contrary to alpine glaciers, hot but dry episodes such as those observed during the summer of 2003 in a large part of western Europe, will therefore have only reduced effects on the mass balance of cave ice fillings. This point is well supported by observations carried out in numerous ice caves in the Jura Mountains. Figure 2 illustrates ice fluctuations measured in Monlési ice cave over a three-year observation period. Data reflect the ice elevation at a station located on top of the ice filling. Seasonal variations are shown and outline a negative trend of annual ice mass balances. In comparison to the two previous years, no significant increase of melting rates could be observed during the summer of 2003.

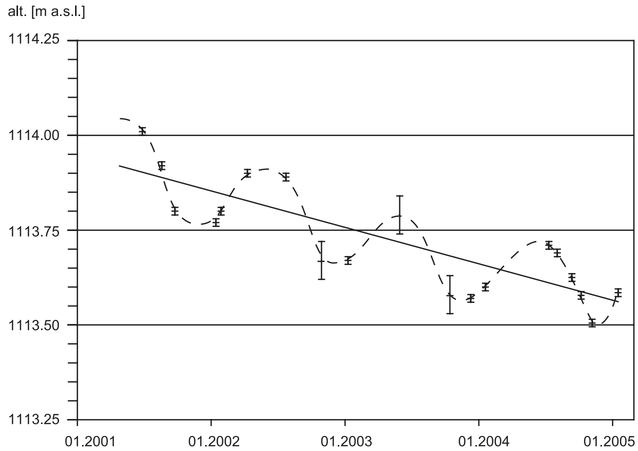

Fig. 2 Ice fluctuations measured in Monlési ice cave between 2001 and 2004. Although a negative trend of the ice mass balance can be observed, an exceptionally hot period in the summer of 2003 had no significant impact on the ice volume.

Ice characteristics	**Estimated ice volume** (2003)	**Type of ice cave** (after Luetscher & Jeannin 2004)	**Remarks**
congelation ice	~6000	statodynamic with congelation ice and firn	snow at the base of the entrance
firn	~3000	static with firn	stratified ice front of 20 x 7 m
congelation ice	~500	static with congelation ice	
firn	~200	static with firn	
firn	~160	static with firn	
firn & congelation ice	45-60	static with firn and congelation ice	
firn & congelation ice	~50	static with firn and congelation ice	
firn	<50	static with firn	
firn	<50	static with firn	
firn	<50	static with firn	
firn	~40	static with firn	
firn	~20	static with firn	
firn	~20	static with firn	
firn	~15	static with firn	
firn	~10	static with firn	
firn	~10	static with firn	
firn & congelation ice	10	dynamic with firn and congelation ice	
firn & congelation ice	<10	static with firn and congelation ice	
firn	~9	static with firn	
firn	<5	static with firn	
congelation ice	<1	static with congelation ice and firn	
firn	n.d.	static with firn	~15 m^3 in 1989
firn	n.d.	static with firn	~20 m^3 in 1981
firn	n.d.	static with firn	~700 m^3 in 1978
firn & congelation ice	n.d.	static with firn and congelation ice	

4.2.2
Mass turnover due to melting at the bottom of the ice mass

At a certain depth below ice caves, the temperature of the karst massif is positive. Therefore, a heat flux from the karst towards the ice cave is expected. This heat flux produces a certain amount of melting at the bottom of the ice mass, causing a constant mass turnover of the cave ice. Due to the heterogeneity of karst networks, this heat flux changes from one place to the other and the turnover duration should therefore be significantly different in each ice cave. Proxy records from Monlési ice cave show that this turnover can be relatively fast. A broken tile, caught in an ice strata about 10 m below the ice surface, has been manufactured between 1874-1916 (pers. comm. B. Boschung, SPMS-Neuchâtel). This observation is backed up by metallic objects found more or less at the same depth. It also correlates well with a three year topometric survey using reference points marked in the ice mass (Luetscher, *in press*). Vertical displacement of about 10 cm/year have been measured. The complete turnover of the ice mass in this cave requires therefore about one century, which seems consistent with the number of ice layers observed. In the St-Livres ice cave, photographic documents from the last 50 years seem to indicate a more reduced mass turnover. This estimation, confirmed by indirect dating provided by dendrochronological analyses of tree trunks caught in the ice filling (Schlatter et al., 2003), leads to an age of about 150 years for the upper layers. A ^{14}C dating of a wood sample caught in the lower layers gave an age of 970 ± 45 BP (ETH-28592). Although these results must be validated by further samples, they seem to confirm the possible presence of several hundred years old cave ice in the Jura Mountains.

4.3 Fluctuations of cave ice mass balance

Due to the difficulty of access to most of these caves, observations of ice volumes are scarce and mainly recent. The first systematic descriptions of ice caves were carried out during the nineteenth century by different naturalists (e.g. Thury, 1861; Browne, 1865; Girardot and Trouillet, 1885; Magnin, 1900) but cave maps are available for the most important ice caves in the Jura Mountains only since the nineteen fifties. However, following the improvement of surveying and mapping methods, some of these caves were mapped several times within a few decades, allowing a good overview of ice volume fluctuations (fig. 3).
Furthermore, several pictures taken during the 20th century (mostly for souvenir purposes) could be collected. Therefore, photographic documents are available of the last few decades for major ice caves of the Jura Mountains. The collection and analysis of these documents provides a precious historical archive for a better assessment of subsurface ice mass evolution. Brulhart (2001) first attempted such a systematic reconstruction for the Glacière de St-George and

Tab. 1b

	Name	E	N	Z	Number of entrances	Last observed perennial ice
1	Baume de la Roguine	6°38’25.6’’	46°52’38.5’’	1230	1	~1984
2	Baume Est du Croue	6°8’ 5.7’’	46°29’52.4’’	1490	1	~1992
3	Baume Ouest du Petit-Pré	5°18’2.0’’	46°34’32.4’’	1460	1	~1995
4	Creux à Neige du Chalet Neuf	6°15’58.8’’	46°33’13.4’’	1460	1	n.d.
5	Glacière Le grand Abergement	n.d.	n.d.	n.d.	1	~begin 20th century
6	Glacière à Tissot	6°19’38.2’’	46°35’22.0’’	1380	1	n.d.
7	Glacière d’Arc-sous-Cicon	6°23’59.2’’	47°1’42.5’’	1070	1	~begin 20th century
8	Glacière de Gonnefay	6°44’56.5’’	42°2’14.6’’	900	1	~begin 20th century
9	Glacière du Bois Claude	6°43’10.4’’	47°7’59.9’’	920	1	~begin 20th century
10	Glacière de la Genolière	6°7’28.2’’	46°27’56.3’’	1340	2	~begin 20th century
11	Glacière de Vergy	n.d.	n.d.	n.d.	1	~begin 20th century
12	Glacière des Baumes	6°29’6.9’’	46°53’15.4’’	1178	2	~1965
13	Glacière des Granges du Roi	6°35’46.5’’	47°14’16.5’’	720	1	~begin 20th century
14	Glacière du Bois du Roi	6°33’39.2’’	47°12’1.8’’	615	1	~1900
15	Glacière du Luisans (1)	6°33’19.8’’	47°5’25.5’’	990	1	n.d.
16	Glacière du Luisans (2)	6°33’19.8’’	47°5’25.5’’	990	1	n.d.
17	Glacière Hauteville	n.d.	n.d.	n.d.	1	~begin 20th century
18	Glacière sud (1) du Mont-Tendre	6°18’57’’	46°35’34.5’’	1565	1	~1963
19	Gouffre de Vers-chez-Amiet	6°42’11.8’’	46°54’45.2’’	1295	1	~2002
20	Grand Creux des Glaces	7°9’5.3’’	47°13’6.8’’	995	1	n.d.
21	Grotte de la Glacière	6°21’43.3’’	47°14’78.7’’	525	1	~begin 20th century
22	Neigières de Gilley	n.d.	n.d.	n.d.	1	~begin 20th century
23	Petit Creux des Glaces	7°9’3.0’’	47°13’6.0’’	998	1	n.d.

concluded that a general decrease of the ice in this cave was observable during the 20th century. The same approach has been attempted here based on all available data from the Jura mountains' ice caves. The data processing was confronted to highly heterogeneous material related to their origin, the type of documents and discontinuous time series.

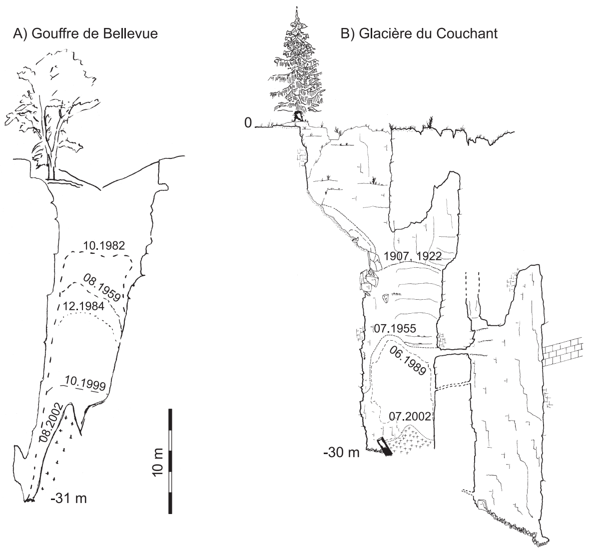

Fig. 3 Firn extension in two static ice caves. Both vertical cross sections show the evolution of the observed ice filling during the last decades of the 20th century. Mass balance seems to have been quite stable until the end of the eighties but a strong negative trend has been observed since 1989.

Remarks

abundant cryoclasts, soutirage attributed to melting ice
perennial firn until mid eighties
abundant cryoclasts, presence of seasonnal ice
presence of seasonnal ice
mentioned in Magnin 1900
presence of seasonnal ice
mentioned in Magnin 1900
mentioned in Magnin 1900
mentioned in GIPEK 2004
presence of seasonnal ice
mentioned in Magnin 1900
cryoclasts, sorted sediments
seasonnal ice accumulations
mentioned in Magnin 1900
5-6 m of firn in 1900
mentioned in Magnin 1900
mentioned in Magnin 1900

no more accessible
first mentioned in 1586; ice is artificially maintained
mentioned in Magnin 1900

Figure 4 shows a reconstruction of the evolution of ice volumes in five of the documented caves. Historical data used for volume estimations come either from topographic maps, picture comparison or inscriptions found in places in the cave which are no longer accessible today, etc. This compilation confirms a general decrease of ice volumes over the 20th century in most ice caves in the Jura Mountains. However, the trend is all but consistent over the entire observation period. If cave ice mass balances seem to have been almost in equilibrium up to the end of the nineteen eighties, a noticeable decrease of ice volumes can be observed for the last fifteen years. This negative trend is well evidenced in vertical shafts where reduced winter snow accumulation are no longer able to compensate the annual melting (for instance, Gouffre de Bellevue, fig. 3). Consequently, ice in many of the smallest ice caves, has already completely disappeared (table 1b).

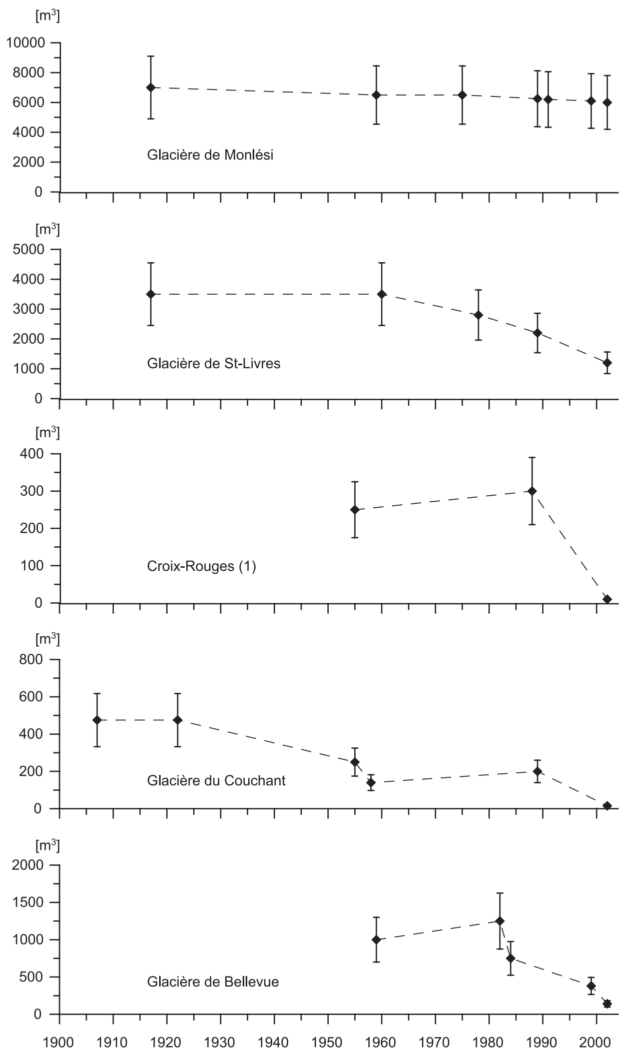

Fig. 4 Reconstructed cave ice volume fluctuations on selected sites in the Jura Mountains. Mass balance estimates outline a strong decrease of cave ice during the last decade. Error bars reflect the estimated inaccuracy (30%) attributed to measurements and seasonal variations of ice volume.

However, various periglacial signatures might be well preserved in cave sediments, among which cryoclasts remain probably the most common features (e.g. Kempe and Rosendahl, 2003). These can be observed in most cave entrances but could also be described as low as one hundred meters below surface (for instance in Glacière de Druchaux / Berolles, VD, table 1a). Furthermore, periglacial processes may also have affected cave sediments leading to ground cryoturbation features like stone circles, sorted polygons, stripes and clay hummocks (Mihevc, 2003). In fact, such evidences have been found in a few caves of the Jura Mountains. Figure 5 illustrates some of the sorted patterns found in the Glacière des Baumes (Les Verrières / NE, table 1b). Since perennial ice fillings disappeared from this cave at the end of the nineteen sixties, such cave sediments still represent an excellent indicator for the presence of freezing and thawing cycles (Kessler and Werner 2003). Owing to the reduced erosion occurring in fossil caves, such patterns can be preserved over several thousands of years. Therefore, the identification of periglacial patterns in low altitude cave sediments constitutes a reliable marker for Quaternary winter climate fluctuations and their investigation could complete profitably more traditional investigations of subsurface sediments.

Fig. 5 Sorted stone patterns (scale = 30 cm) observed in the glacière des Baulmes (Verrières / NE). The disappearance of numerous low altitude ice fillings lead to distinct periglacial signatures in several caves. These can be used for investigating winter paleoclimate conditions.

5 | Interpretations

5.1 Relationship between outside climate and subsurface ice occurrences

Mass variations (Δm_{ice}) of subsurface ice accumulations result from a difference between seasonal ice accumulation (m_{new_ice}) and annual melting (m_{melted_ice}):

$$\Delta m_{ice} = m_{new\_ice} - m_{melted\_ice} \quad (1)$$

Hence, a negative mass balance can either result from a decrease of the seasonal ice accumulation or from an increase of melting rates.

In cold air traps, the energy supply leading to ice melting results mainly from heat exchanges with the surrounding rock (i.e. ground heat flux) and from summer infiltration water (Luetscher et al., 2003). Thanks to the inertia induced by the heat capacity of the rock (Badino, 1995), ground heat fluxes can be considered as stable over several years. Since long time-series do not show any significant variation in summer precipitation regimes during the 20^{th} century (Bader, 2002), multi-annual melting rates of cave ice can be considered as more or less constant. This leads to the postulate that the observed ice mass decrease is mainly controlled by the annual accumulation of ice and not by the annual melting rate.

The formation of endogenic cave ice relies on the cooling process of the cave, which is dominated by heat exchanges induced by winter air circulations (e.g. Luetscher and Jeannin submitted a). Since air circulations in ice caves are related to density differences between inside and outside air (e.g. Lismonde, 2002), the heat lost by the system, i.e. responsible of the freezing, depends on the time integration of temperatures below 0 °C, weighted by the air flow circulating through the cave. Thus, one can hypothesize that subsurface ice mass balances mainly reflect specific variations of exterior winter climatic conditions.

Figure 6a illustrates air temperatures measured at the station of Chaumont during the 20^{th} century. A 20-year moving average plot outlines a general trend of about 1 °C/100 years ($r^2 = 0.77$). This tendency is more or less equally distributed between summer and winter semesters (fig. 6b&c). Regarding ice caves, a significant change occurred in the nineteen forties, when mean winter temperatures, which were mostly below 0 °C at this altitude, rised up above freezing point. However, between 1960 and 1988 this trend reversed, with winter temperatures reaching slightly below 0 °C in the late seventies. Since 1989, winters have become clearly milder.

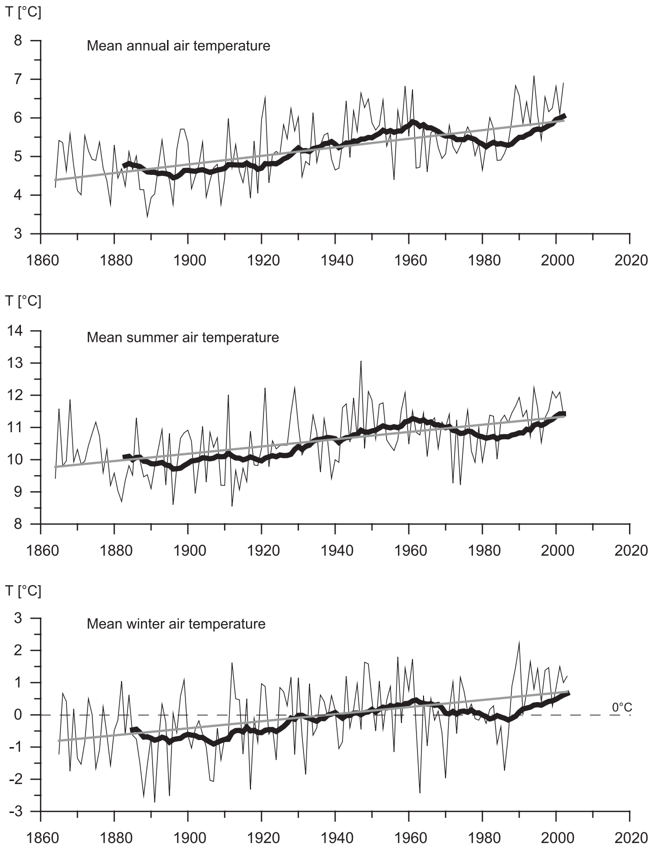

Fig. 6 Mean air temperature evolution measured in the Jura Mountains (Meteoswiss data). A twenty-year moving average (bold line) outlines a general trend of about 1 °C/100 years at the station of Chaumont (r^2: 0.77). This trend is more or less equally distributed between summer and winter semesters. But since the nineteen forties, mean winter temperatures, which were usually below 0 °C at this altitude, became mostly warmer than the freezing point.

Consequently, and especially since 1989, freezing days (days where **D**aily **M**ean **A**ir **T**emperature is below 0 °C) became less frequent and the annual freezing index («FI»: integer of DMAT<0 °C) significantly decreased. This trend is well supported by figure 7a & b, which outlines a reduction of the FI of about 40% (i.e. ~130 °C*day).
Measurements performed in Monlési ice cave during the winter season show a mean temperature difference between the cave air and the exterior atmosphere of 2 °C, what corresponds to an air flow of about 3 m^3s^{-1} (Luetscher and Jeannin, submitted a). From this value, one can compute the impact of a reduced freezing index: the potential decrease of ice freezing is estimated at about 100 m^3 per year in Monlési ice cave.
Furthermore, the significant decrease of snow precipitations observed during the same period (fig. 7c) induced a deficit of latent heat in form of snow in low altitude cave entrances. Also, the absence of a thick snow cover at the beginning of the spring snow melt reduces the water infiltration which can potentially freeze shortly after a cold season. As a result, neither winter cooling nor water presence were sufficient for providing equilibrated mass balances during the last decade of the 20th century.

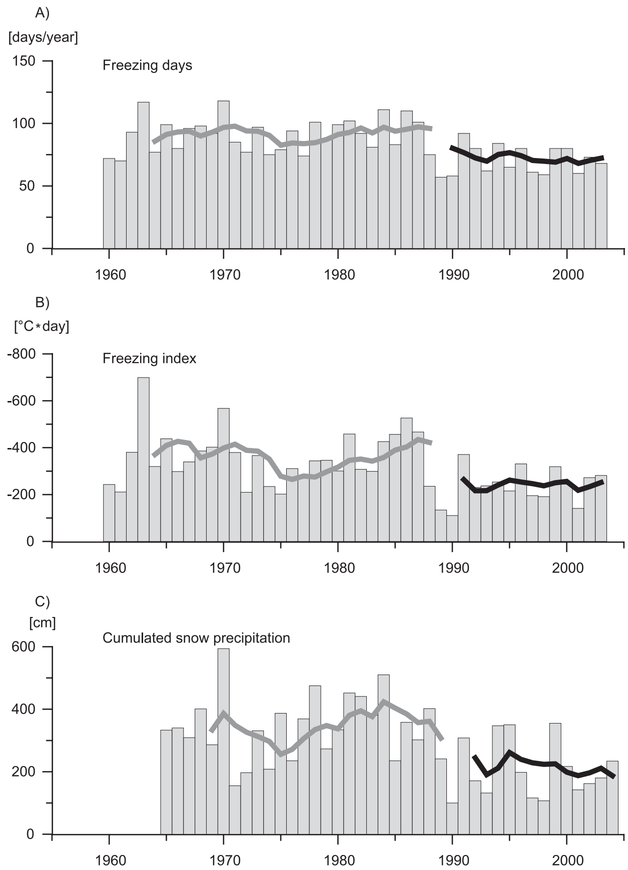

Fig. 7 Time series of winter climatic parameters in the Jura Mountains. A five-year moving average underlines the major change observed at the end of the nineteen eighties (bold lines). A) Number of freezing days (Daily MAT < 0 °C); B) Freezing index [°C * day] and C) Cumulated snow precipitations [cm]. Daily records, provided for the station of Chaumont by meteoswiss since 1959, demonstrate a significant change in winter climate since 1989: milder temperatures are observed simultaneously with decrease of snow accumulation.

So far, continuous observations of ice mass balance in caves are too scarce. Moreover, subsurface heat exchanges (i.e. cave ice mass balances) are directly controlled by each specific cave geometry and this aspect was not integrated in the analysis yet. However, the present study points out the main factors controlling ice formation as well as the general trend of ice mass balance in caves of the Jura Mountains. This data is being used to compute physically based models of some selected sites, which will significantly increase our understanding of ice caves.

6 | Discussion

The study of processes at the origin of cave ice led us to infer that mass balances reflect the winter climate fluctuations. Summer air temperatures appear to play only a minor role on the annual energy balance.

Literature research and detailed field investigations enable to list 25 caves presenting a perennial ice filling in the Jura Mountains. The observation of their dynamics confirms that cave ice does not have to be considered as a relict of the last ice age. The ice mass balance is controlled by the recent and present day climatic context and is mainly related to the winter temperature-precipitation regime. Reconstructed data show that most of the ice fillings present a marked negative mass balance since 1989. This trend is well correlated to recent climatic changes outlining a decrease of low altitude snow precipitations with a simultaneous to a general increase of mean winter air temperatures. On the other hand, exceptionally high outside air temperatures, like those observed in the summer of 2003, induced no significant increase of melting rates in the investigated caves.

These latter observations indicate that all (but one) ice cave in the Jura Mountains behave as cold air traps. Since most cave ice results from firn accumulations the deficit in snow precipitation observed during the investigated time window significantly reduced the annual ice formation. Concurrently, at altitudes where the mean winter temperature is close to 0 °C, a slight temperature increase reduces significantly the formation of congelation ice. Consequently, while heat exchanges leading to melting processes (e.g. ground heat flux, water infiltration) remained almost stable, the deficit of annual ice accumulation lead to the observed negative mass balances.

It could be demonstrated that, due to a relatively rapid mass turnover, the age of cave ice in the Jura Mountains probably does not exceed several hundred years. Therefore paleoclimatic records in cave ice layers can document only short time periods compared to other cave sediments. Nevertheless, cave ice constitutes a remarkable object for the investigation of recent atmospheric changes and the potential represented by ice chemistry (incl. oxygen isotopes) can still be exploited. But due to seasonal melting processes at the ice surface, remobilization of glaciochemical components occurs and might significantly disturb the original stratigraphy. Since ice caves represent a new topic for the study of Holocene paleoclimate in regions missing useful paleoclimatic archives, one of the first challenges consists in providing absolute datings of the subsurface ice fillings.

If, in the Jura Mountains, the presence of ice caves has been confirmed by historical data since the XVIth century in the biggest caves (e.g. AEN, 1554; Poissenot, 1585), recent variations of the equilibrium altitude of ice caves induced the disappearance of several low altitude subsurface ice patterns. However, evidence of past cryogenic processes and/or a presence of former ice fillings could be outlined in many caves, even though they are no longer located in any periglacial context (e.g. Pissart et al., 1988; Mihevc, 2003). Ground cryoturbation features and/ or indications of former ice fillings could be described in several caves in the Jura Mountains as low as 525 m a.s.l. Since during the investigated time period no ice caves presented an equilibrated mass balance, it is supposed that the Equilibrium Line Altitude (ELA) of ice caves moved up several hundred meters compared to the colder periods of the Holocene. Given that cave sediments can be well preserved from external influences (for instance erosion, weathering,...), signatures of cryogenic processes can remain for centuries or even millennia. Therefore, they provide indications of former ice deposits in regions where climatic conditions do no longer show freezing temperatures. The study of these cave sediments may thus represent a new investigation field for identifying cold episodes in temperate low altitude regions and constitute a useful indicator of past climatic conditions (for instance of continental climate or glacial periods).

This study outlined the importance of winter temperature and precipitation regimes for the conservation of perennial cave ice. With an expected decrease of low altitude snow precipitations (e.g. Scherrer et al., 2004), numerous subsurface firn accumulation will disappear. However, the alternation of cold-dry and wet-mild episodes (rain or melting snow) constitute ideal conditions for the formation and preservation of subsurface congelation ice. Such a situation might occur even in a warming climate context. It must be emphasized that at present no detailed climatic model is able to predict the evolution of the winter temperature and precipitation regimes with a sufficient precision to truly estimate what is the future of ice caves in the Jura Mountains. In any case, a detailed monitoring of mass balances in selected ice caves is recommended.

7 | Acknowledgements

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Is dating of mid-latitude/low-altitude cave ice possible?

– a case study from the Swiss Jura Mountains – article in preparation

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Abstract

Investigations conducted in the Jura Mountains demonstrate that dating of mid-latitude/low-altitude cave ice is possible using a multi-parametric approach, involving stratigraphic markers and radiogenic isotopes. Although melting processes are frequent, radiogenic dating with U-series isotopes is reliable if input (i.e. radium, radon) is controlled. Results from two selected study sites in the Jura Mountains (Switzerland), Monlési and St-Livres ice caves, emphasize that accessible time periods can vary significantly from one site to the other. Approximately 120 and 1200 years old, respectively, these perennial cave ice accumulations are excellent objects for explorative studies of recent winter climatic conditions in mid-latitude/low-altitude environments.

Key words: ice cave, dating, isotope, environmental record.

1 | Introduction

Mid-latitude glaciers are natural archives, well suited for studying past environmental and climatic conditions (e.g. Cecil et al., 2004a). Studies have focussed on cold, high-alpine glaciers, where meltwater formation and percolation, which could destroy the glaciochemical signature, are negligible. In the Alps, such glaciers are found at altitudes above 4000 m a.s.l. Consequently, suitable glacier archives are rare and the accessible paleoclimate information is spatially very limited. In order to enlarge the spatial coverage, the potential of temperate glaciers to preserve valuable climatic records has recently been investigated. Whereas trace species in the firn part of temperate glaciers seem vulnerable to meltwater percolation, firn and ice matrix parameters such as the stable isotope ratios might be preserved (Eichler et al., 2001; Pohjola et al., 2002; Schotterer et al., 2004).

In this context, subsurface ice accumulations represent an innovative research topic which was frequently overlooked until recently. Sometimes located at altitudes well below the 0 °C-isotherm, ice caves could provide valuable paleoclimatic data in regions where such records are often missing (Perroux, 2001). By studying the processes controlling the presence of cave ice, Luetscher et al. (2005) demonstrated a close relation between subsurface ice accumulation and the winter temperature and precipitation regime. Since it could be shown that summer climatic conditions play only a negligible role on the annual mass balance, results expected from glaciochemical investigations on cave ice mainly reflect the evolution of winter climate. Such proxy data for winter climate are particularly rare, as many paleoclimate archives preserve mainly summer conditions, with tree rings being only one example. The preliminary cave ice data instigated detailed glaciochemical investigations in several European and North American ice caves (e.g. Cecil et al., 2004b; Turri et al., 2003; Holmlund et al., in press), but cave ice records can only be correctly interpreted when based on a reliable dating. Most authors agree that low-altitude cave ice is a result of current accumulation processes, but little information is available on the accessible time period in ice caves. The main dating tool applied in ice core studies is the counting of annual layers of one or more seasonally varying parameters. Seasonal parameters include the stable isotope ratios $\delta^{18}O$ or δD, a bulk parameter such as dust or the acidity content, the visual stratigraphy, and the concentration of a chemical tracer, such as NH_4^+, Ca^{2+} or NO_3^-. One of the prerequisites for annual layer counting is the preservation of snow during all seasons, which is not given for subsurface ice accumulations. Nevertheless, since the ice accumulations are often located in the proximity of cave entrances, seasonal organic deposits (soil, wood, etc.) can be frequent. These deposits are a good indicator of cave ice accumulation rates and may provide useful dating approximations. A detailed analysis of this litter material (for instance palynology, dendrochronology, ^{14}C) represents a valuable approach to the rough characterization of the ice chrono-stratigraphy. Actually, sparse ^{14}C-datings of organic material enclosed in massive subsurface ice accumulations suggests that thousand-year-old ice fillings could still subsist (e.g. Pavuza & Spötl, 1999; Achleitner, 1995; Fanuel, 1993; Schroeder, 1977; Lauriol & Clark, 1993). However, due to frequent melting processes, significant gaps are expected between the different layers. The present study aims to test several other methods, normally applied to the dating of alpine ice cores, with special focus on nuclear dating using radioactive isotopes. For the case study, two ice caves in the Swiss Jura Mountains were selected, representing different ice formation mechanisms. The ice in the Monlési cave is formed mainly by refreezing of snow meltwater (congelation ice), whereas firn accumulation represents the major ice formation process in the St-Livres cave. Several of these dating methods were applied for the first time to subsurface environments.

2 | Experimental

2.1 Site selection

The Jura Mountains form an approximately 400 km wide arc between the Savoie (France) in the southwest and the Black Forest (Germany) in the northeast. The inner part of this northwest arcuate range, mostly located in Switzerland, is characterized by a succession of crests and valleys ranging between 1000 and 1500 m a.s.l., with the highest peaks reaching about 1700 m a.s.l. Mean annual air temperature measured at these altitudes is between 3.5 and 6.5 °C. Although this mountain range does not belong to previously recognized permafrost areas (Keller et al., 1998), 24 perennial cave ice fillings were identified in the Jura Mountains (Luetscher et al., 2005). Heat exchange allowing the conservation of subsurface ice accumulations is mostly initiated by gravitational effects of the cold exterior winter air. Two sites were selected for further glaciochemical investigations: «Glacière de Monlési» (6°35'4''/46°56'18'', 1135 m a.s.l.) and «Glacière de St-Livres» (6°17'50''/46°33'47'', 1359 m a.s.l.). The main criteria for the selection of these two ice caves were the large amounts of massive ice, visible and accessible ice stratification and the presence of numerous datable clasts. For both sites an extensive documentation exists, compiled by speleologists since the mid nineteen-fifties (e.g. Stettler & Monard, 1960; Stettler, 1971; Luetscher & Wenger, 2002; Dutruit, 1991; Audétat et al., 2002). Located in two distinct geographical regions (Fig. 1), these caves also differ from each other in the nature of their ice content.

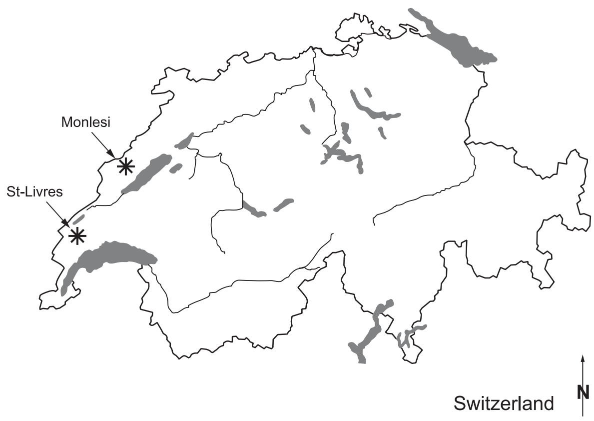


Fig. 1 Geographical situation of Monlési and St-Livres ice caves. Both caves are located in a climatic context where the MAAT is well above 0 °C.

The filling of Monlési ice cave results mainly from the accumulation of annual deposits of congelation ice. The cave opens with three entrance shafts leading at -20 m to a large room of about 20 x 40 x 15 m. Luetscher & Wenger (2002) estimated the ice volume at 6000 m^3 where morphological evidence suggests a maximal ice thickness of about 12-15 m (Fig. 2a). Ice crystallization occurs preferentially during spring, when exterior snow melting enables water infiltrations. Measured values show a seasonal ice accumulation rate ranging from 10 to 30 cm per year, though a major fraction melts again during the course of the year, leading to strata of about 5-10 cm thickness (Luetscher, 2004). The low permeability of this massive ice suggests a reduced risk of remobilization due to percolating meltwater.

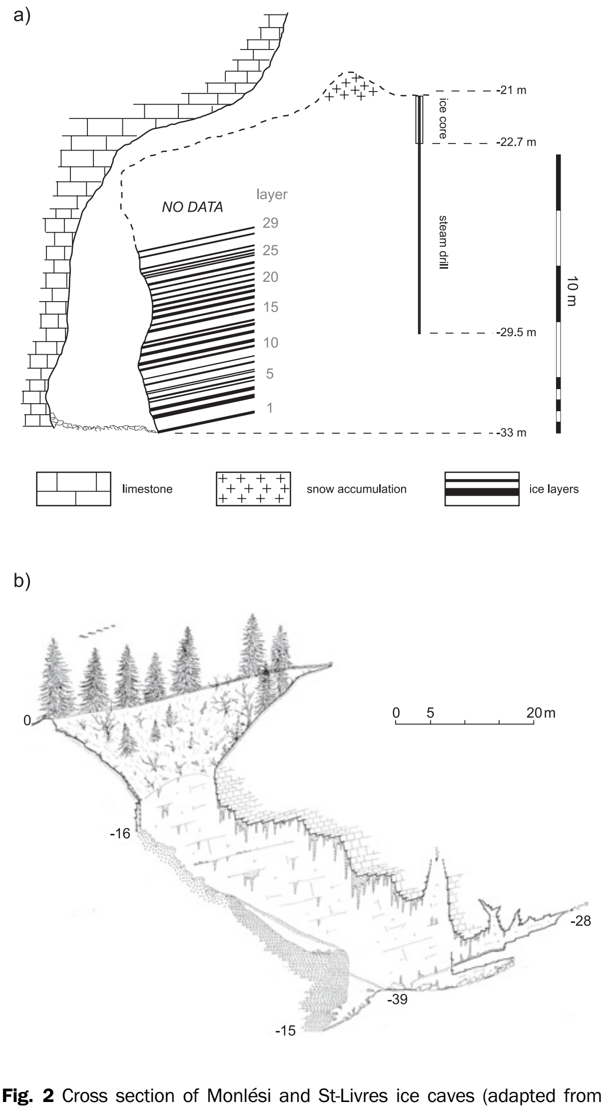


Fig. 2 Cross section of Monlési and St-Livres ice caves (adapted from Dutruit 1991). Owing to the presence of frequent clastic material, ice coring could not be performed on the total ice thickness.

Contrary to the Monlési ice cave, firn accumulation represents the major process at the origin of the St-Livres ice cave. A large collapsed doline (Ø: ~20 m) constitutes the unique entrance of this cave, and leads to the deepest part of the cavity at -45 m. The ice filling, with an estimated volume of 1200 m^3, occupies the base of the entrance shaft (Fig. 2b). It is formed essentially from the diagenesis of snow accumulated during winter, but local refreezing processes of infiltration water contribute also to the actual ice mass. Owing to the large cave entrance located vertically above the ice mass, major organic deposits are observed at the ice surface.

2.2 Sampling

Since stratigraphical observations showed that numerous distinct ice layers are present in both caves, explorative steam drilling (*Heucke ice drill system*) was performed to determine the total accessible ice thickness. Although the presence of abundant clastic sediments limited these soundings, an 8.5 m deep borehole was drilled in the Monlési ice cave. Using Pt-100 thermistors, ice temperatures were monitored at different depths (-1 m; -2 m; -4 m; -5 m and -8 m) from November 2002 to October 2003. Ice core drilling with the small light-weight coring-system «Felics» (Ginot et al., 2002) proved to be difficult due to ice temperatures close to the melting point and the presence of clastic sediments. Nevertheless, ice chips with a volume of a few cubic centimetres were sampled down to a depth of -1.7 m. Some core samples were composed of clear ice while others contained organic and sedimentary debris. Fifteen additional ice samples were collected manually from the accessible part of the ice front (-5.5 to -12 m stratigraphical depth) using a cordless drill hammer equipped with a hole saw. To avoid contamination by flowing meltwater on the ice surface, the outermost centimetres were removed before sampling. The ice samples (Ø 8 cm, 5 cm thick) were packed into polyethylene tubes and transported in dry ice to a cold room kept at -25 °C.
In addition, water samples were collected manually at the main water inlets. For comparison, data were collected from the nearby precipitation station (La Brévine, Neuchâtel) of the Swiss National Network for the Observation of Isotopes in the Water Cycle (NISOT), where tritium, $\delta^{18}O$ and δD have been measured in monthly composite samples since 1994 (Schürch et al. 2003).

2.3 Analytical procedures

Carbon-14 analyses were performed at the AMS-laboratory of the ETH-Zurich by measurement of the $^{14}C/^{12}C$ ratio. Wood samples underwent pre-treatment in a soxhlet apparatus, involving baths of hexane, acetone and ethanol, followed by the standard acid-alkali-acid treatment. The procedure described by Vogel et al. (1984) was applied for graphitization. Calibration was performed using the program CalibETH (Niklaus et al., 1992).
Tritium content (sample volume 10 ml) was determined by direct β^- measurements in a liquid-scintillation spectrometer (Schotterer et al., 1998) at the Physics Institute, University of Bern.

The detection limit on a 2σ base (σ=standard deviation) is 1.6 TU. The ^{210}Pb activity concentration (sample volume 200 ml) was indirectly determined from the activity of its granddaughter nuclide ^{210}Po, electrolytically deposited on Ag plates. The ^{210}Po activity was determined by measuring its a decay at an energy of 5.3 MeV (Gäggeler et al., 1983). Radon in water samples (20 ml) was determined by liquid scintillation counting (Canberra Packard Tri-Carb 2250CA) at the Centre of Hydrogeology of Neuchâtel (CHYN lab). Cave air samples were collected by passing approximately 2 l of air through 180 ml Lucas-cells. The Lucas-cells then were measured at the CHYN-lab (RDA-200, Scintrex). ^{238}U, ^{226}Rn and ^{210}Pb in rock, soil and litter samples (for instance, organic material) were determined by γ-spectrometry (HPGe well-type detector, CHYN-lab). $\delta^{18}O$ analyses of cave ice were carried out at the Paul Scherrer Institute by pyrolisis of the liquid sample at 1450 °C in a glassy carbon reactor to produce carbon monoxide (CO). The $C^{18}O/C^{16}O$ ratio of the gas was measured using an isotope ratio mass spectrometer (Delta Plus XL, Finnigan MAT). Results are reported to the Vienna Standard Mean Ocean Water (VSMOW, Baertschi, 1976):

$$\delta^{18}O\ [^0/_{00}] = [(R_{sample}-R_{standard})/R_{standard}]*1000$$

Where R is the ratio $^{18}O/^{16}O$ and $R_{standard}$=(2005.20±0.45) * 10^{-6}

Major ions (K^+, Na^+, Mg^{2+}, Ca^{2+}, SO_4^{2-}, Cl^-, NO_3^-) were identified after a 0.45 mm filtration at the Centre of Hydrogeology, UNINE, by ion chromatography (Dionex model DX-120, with a column No. AS14A for anions, and No. CS12A for cations).

3 | Results & Discussion

3.1 Ice characterization: temperature & lithologies

Borehole temperatures recorded in Monlési cave ice from November 2002 to October 2003 (Fig. 3) showed seasonal temperature oscillations along the entire drill hole. However, data recorded at -8 m suggest a temperate ice filling affected from all sides by melting processes. Regardless of smaller fractures and clasts encountered during the drilling, 2D thermal modelling (Luetscher et al., in press) suggests that the ice volume is homogeneous at the macro scale. Hence, it is assumed that percolating melt-water is negligible, which is supported by a low porosity of the observed cave ice (density ~920 kgm^{-3}).

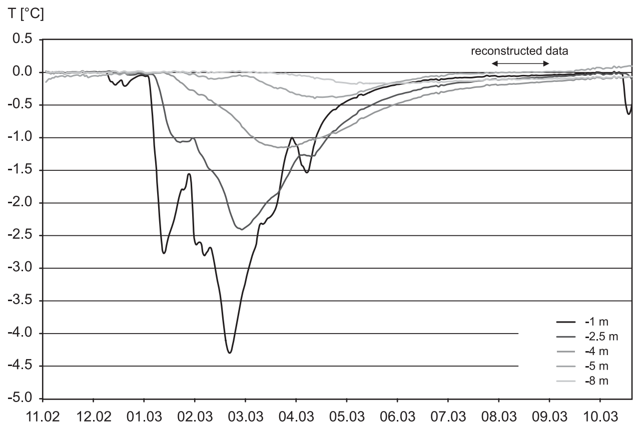


Fig. 3 Daily mean temperature recorded at different depths in the ice filling of Monlési cave during the annual cycle 2002-2003. Seasonal fluctuations are observed over the entire filling. Data suggests a temperate ice body.

Although chemical differentiation between congelation ice and sedimentary ice is possible based on the content of dissolved carbonate (e.g. Shumskii, p. 96; Table 1), the distinction between both lithologies relied mostly on genetic processes often determined by simple field observations. Observations performed at St-Livres ice cave concluded that congelation ice consists of thin-layered orientated centimetre-size ice crystals attributed to individual freezing events. These congelation ice layers are well distinguishable in situ, because of their strong absorption of light. Conversely, firn ice has an anisotropic structure constituted of coarse grained elements. Firn layers are mostly parallel to the substratum and contain frequent organic clasts in their upper section. These layers are assumed to represent seasonal deposits and thus, are interpreted as annual strata. Nevertheless, complex sedimentary profiles were observed showing an interstratification between both lithologies.

3.2 Dating by determining mass turnover rates

A three year topometric survey conducted at Monlési ice cave (Luetscher, 2004) revealed a basal melting rate of

Tab. 1 Major ions and electrical conductivity of ice samples from Monlesi and St-Livres ice cave.

Cave	Sample	Na^+ [mg/l]	NH_4^+ [mg/l]	K^+ [mg/l]	Mg^{2+} [mg/l]	Ca^{2+} [mg/l]	K [μS/cm]
Monlési	firn	0.58	0.52	0.97	0.24	9.35	56
Monlési	flowstone	0.09	0.04	0.05	0.07	6.03	30
Monlési	sample # 1	0.30	0.10	0.18	0.43	11.56	63
Monlési	sample # 2	0.31	0.42	0.22	0.19	8.95	43
St-Livres	firn	0.18	0.15	0.14	0.27	4.37	25
St-Livres	flowstone	0.21	0.09	0.12	0.41	8.19	45
St-Livres	sample # 3	1.19	0.26	0.58	0.38	8.85	n.d.
St-Livres	sample # 4	1.53	0.21	0.57	0.31	7.60	n.d.

about 10 cm·year^{-1}. Assuming a constant basal heat-flow and a total ice thickness of 12 m, this value suggests a complete mass turnover about every 120 years.
Although this method is reliable for estimating the age of the cave ice, it requires long-term observations. In cases of lower mass-turnover rates, the displacement induced by the basal melting might be too small to be observed. Such low mass-turnover rates are suggested by photographic documentation of St-Livres ice cave during the last 30 years. These pictures indicate that decadal basal melting rates are almost insignificant in this cave. Although older documents are available, they are not accurate enough to allow a better assessment of this turnover rate.

3.3 Dating of clastic material

The near proximity of cave entrances leads to frequent organic deposits on the cave ice surface. This external litter material is often associated with catastrophic events (for instance small-scale landslides or falling trees) during the summer/autumn season. Occurring after the maximal crystallization phase of the cave ice, this material constitutes a good marker of the annual periodicity of ice formation. Measurements demonstrated that the exogenic origin of this litter material can often be determined by high ^{40}K and ^{137}Cs activities in the samples (see also Table 5).
Conversely, endogenic clasts are frequently related to frost shattering of the cave walls and to carbonate precipitates formed by the segregation of solutes or insoluble residues during freezing of percolating drip-water. Measurements performed in Monlési ice cave indicate that the dust material in cave ice varies between less than 0.02 gkg^{-1} and more than 2 gkg^{-1}. Recent studies demonstrated that this cryogenic cave calcite could sometimes be dated by U-series (Zak et al., 2004) or ^{14}C (Lauriol & Clark, 1993).
Stratigraphic surveys performed at both study sites (Fig. 4) confirmed the presence of well distinguishable layers enclosing organic clasts (wood, leafs, bones, etc.) and anthropogenic material (pieces of metal, tiles, etc.).

Fig. 4a View of Monlési ice stratification. The presence of well marked organic layers is attributed to major melting periods (summer season). Photo D. Bolius

Fig. 4b View of St-Livres ice stratification. Photo R. Wenger

A ^{14}C analysis of a twig from Monlési ice cave provided an AMS-^{14}C age of 230 ± 45 years B.P. Even if the temporal variation of the ^{14}C production during the last 300 years does not allow a precise dating, one can deduce that the sample is less than 500 years old (Table 2). This upper limit was confirmed by the dating of a tile manufactured between 1874-1916 (pers. comm. B. Boschung, SPMS-Neuchâtel) which suggests that the maximal age of Monlési cave ice is about 130 years.
Although the presence of industrial nails in the upper half of St-Livres ice cave confirmed the presence of relatively young ice, dendrochronological datings performed on tree trunks (Schlatter et al., 2003) suggested that part of the ice could be older than 250 years. Carbon-14 analyses performed on four samples from different stratigraphic layers (Table 2) indicated maximal ages of 1200 ± 50 years BP, and confirmed the presence of major time gaps in the stratigraphic sequence. Thus the turnover rate in St-Livres is much lower than in Monlési ice cave. The difference is mainly attributed to local ventilation features leading to reduced heat exchanges with the surrounding karst system.

3.4 Nuclear dating of ice samples

Tritium (^3H)

Tritium is a short-lived isotope of hydrogen with a half-time life of 12.32 years (Lucas & Unterweger, 2000) which is

Tab. 2 Carbon-14 analyses performed on wood samples from Monlesi and St-Livres ice caves. A major time gap is set in evidence in St-Livres ice cave.

Cave	Stratigraphical depth	AMS-^{14}C Age [years BP]	^{13}C [‰]	Calibrated age [BC/AD]		
Monlési	-12 m	230 ± 45	-26.9 ± 1.2	AD	1518 – 1595	(11.3%)
				AD	1620 – 1694	(37.6%)
				AD	1726 – 1813	(40.8%)
				AD	1918 – 1949	(9.2%)
St-Livres	-2 m	190 ± 45	-25.1 ± 1.2	AD	1643 – 1707	(23.8%)
				AD	1719 – 1821	(50.5%)
				AD	1827 – 1884	(10.2%)
				AD	1913 – 1950	(15.5%)
St-Livres	-3 m	1040 ± 50	-25.5 ± 1.2	AD	890 – 1056	(90.6%)
				AD	1088 – 1122	(5.8%)
St-Livres	-4 m	970 ± 45	-24.6 ± 1.2	AD	990 – 1163	(98.1%)
				AD	1172 – 1184	(1.9%)
St-Livres	-4.5 m	1200 ± 50	-25.6 ± 1.2	AD	706 – 754	(12.2%)
				AD	757 – 903	(74.1%)
				AD	916 – 963	(11.0%)

produced naturally in the upper atmosphere by cosmic radiation. Natural ^3H levels in precipitation are very low, but large amounts of ^3H were released into the atmosphere by thermonuclear bomb tests, resulting in a major peak observed in precipitation in 1963 (e.g. Clark et al., 1997). Since ^3H is a constituent of the water molecule, it is considered a reliable natural tracer for the identification of recent precipitation events.

Present day tritium levels in precipitation in Switzerland are in the order of a few Bq l^{-1} (FOWG, 2004), but local contamination related to watch industries is frequent in the Jura Mountains. Nevertheless, assuming an ^3H-input activity of ~2 Bq kg^{-1} seems to be consistent with field observations. With a detection limit of 1 TU (=0.118 Bq kg^{-1}), ^3H should still be detectable in 50-year-old ice (4 half-lives). This is especially true for the 1963 peak, where ^3H activity in precipitation reached ~1000 Bq kg^{-1} (i.e. ~90 Bq kg^{-1} in 2005), and which is often used as an ice core dating horizon.

Tab. 3 Tritium analyses of ice samples from Monlési cave (01.09.2004). Low values suggest an age higher than 50 years old for the lower part of the ice mass.

Sample	Depth [m]	^3H [Bq kg^{-1}]
exterior snow		2.4±0.3
#2	-0.1	2.3±0.3
#3	-0.6	2.5±0.3
#4	-0.7	3.3±0.3
#5	-0.8	3.0±0.3
#6	-1	3.8±0.3
#7	-5.5	0.4±0.3
#8	-5.8	0.2±0.3
#10	-6.5	0.1±0.2
#11	-6.8	0.1±0.2
#12	-7.2	0.2±0.3
#13	-7.6	0.6±0.3
#14	-8.5	0.4±0.3
#17	-10	0.2±0.3
#19	-10.6	0.0±0.3
#20	-11.6	0.3±0.3

Tritium content of ice samples from Monlési cave was analysed on two distinct series (Table 3). Five samples were taken from the ice core drilled in the upper layers of the ice filling (0-1 m) while 10 further samples were taken manually from the lower ice layers (5.5-12 m). Tritium values measured in the upper 5 samples are consistent with data of the Swiss national network for the observation of isotopes in the water cycle (NISOT; Schürch et al., 2003). The elevated ^3H levels are chiefly attributed to local contamination induced by the watch industry. Unfortunately, accurate dating of ice deposits is not possible because of the significant reduction in emissions in the last 15 years. However, due to the absence of any outstanding peak, it is reasonably assumed that the samples considered here are modern deposits (i.e. < 20 years). Conversely, analyses of samples taken below a 5.5 m depth do not show any significant ^3H content, suggesting an age of more than 50 years.

Lead-210

Lead-210 (half-life time of 22.3 years) is a natural isotope issued from the disintegration chain of ^{238}U. Its presence in the atmosphere is related to the mother nuclide ^{222}Rn, which emanates continuously from the lithosphere:

$^{238}U \rightarrow {}^{234}Th \rightarrow {}^{234}U \rightarrow {}^{230}Th \rightarrow {}^{226}Ra \rightarrow \mathbf{{}^{222}Rn} \rightarrow \ldots \rightarrow \mathbf{{}^{210}Pb} \rightarrow {}^{210}Po$

Attached to aerosol particles, ^{210}Pb returns to the Earth surface as dry or wet deposition after a residence time in the atmosphere ranging from a few days to a few weeks. The exponential decay of ^{210}Pb has been used successfully to date alpine glacier ice deposits on a century timescale (e.g. Gäggeler et al., 1983). This method also can be applied to temperate glaciers or glacier sites with an irregular deposition of precipitation where most of the conventional dating methods cannot be applied (Von Gunten et al., 1983; Gäggeler et al., 1983).

These characteristics suggest that the ^{210}Pb-method could be appropriate in dating subsurface ice fillings. However, accurate dating is only possible under the following conditions (Gäggeler et al., 1983): (1) the mean ^{210}Pb activity in precipitation has remained constant during the last two centuries; (2) the ^{226}Ra concentrations within the firn/ice samples are negligible; (3) there is no advection of air into the glacier (bearing additional ^{222}Rn) and (4) there is no remobilization of ^{210}Pb (i.e. no cross contamination due to meltwater).

Unfortunately, none of these conditions is perfectly fulfilled. Gäggeler (1995) observed strong seasonal variations of the ^{210}Pb activity at the Jungfraujoch (3450 m a.s.l.) which were attributed to a higher frequency of convective weather types during the summer season. Nevertheless, empirical estimations of a mean annual ^{210}Pb activity were provided for several sites in Switzerland (Von Gunten & Moser, 1993; Von Gunten et al., 1983; Schotterer et al., 1977). Caillet (1999) estimated annual deposition rates at about 170 Bqm^{-2} in the area of the Jura Mountains. These investigations suggest that a constant ^{210}Pb activity in winter precipitation can be assumed.

Condition (2) is correlated with the amount of clastic material present in the ice. Since limestone does not contain significant amounts of ^{226}Ra (e.g. Surbeck & Piller, 1992), contamination primarily comes from the presence of exterior soil and organic material, most frequently found in the area of the cave entrance. Contamination due to advection (3) is almost insignificant in massive congelation ice, but could play a role in porous firn accumulations. Finally, because melting processes are common in most low-altitude ice caves, remobilization of ^{210}Pb is expected at the ice surface and could lead to a differentiated activity. Investigations of Monlési and St-Livres ice caves provide original data for the evaluation of the ^{210}Pb dating method applied to subsurface ice fillings.

Tab. 4 Lead-210 activity of ice samples from Monlési and St-Livres ice caves. Samples from St-Livres ice cave show an anormal high activity which is attributed to the concentrated presence of radon.

Cave	Sample	Depth [m]	^{210} Pb [mBq/kg]	Remarks
Monlési	# 1	-0.1	82.2	clear ice
Monlési	# 7	-5.5	130.4	sediments
Monlési	# 8	-5.8	20.7	
Monlési	# 9	-6.2	113.7	
Monlési	# 10	-6.5	204.6	high sediment content
Monlési	# 11	-6.8	32.7	clear ice
Monlési	# 12	-7.2	17.3	clear ice
Monlési	# 13	-7.6	28.1	clear ice
Monlési	# 14	-8.5	28.9	clear ice
Monlési	# 15	-9	38.7	
Monlési	# 17	-10	49.3	sediments
Monlési	# 18	-10.4	17.4	
Monlési	# 19	-10.6	11.2	clear ice
Monlési	# 20	-11.6	18.5	low sediment content
St-Livres	# 1	-1.5	630.4	firn
St-Livres	# 2	-3	385.5	firn

Table 4 illustrates results of ^{210}Pb analyses performed on 16 ice samples taken at different stratigraphic depths in Monlési and St-Livres ice caves. In Monlési ice cave, data range between 11 and 205 mBq/kg and show no apparent consistency with the observed cave ice stratigraphy. However, it should be noted that elevated ^{210}Pb contents are closely related to the presence of sediments within the ice samples. Despite the few data available, a decreasing trend can be observed with depth if only clear massive congelation ice samples are considered.
As confirmed by the analyses of a composite limestone sample issued from Monlési ice cave (Table 5), the U-series seems to be in equilibrium. Lithoclasts should not be considered as a significant source of contamination because of their small ^{238}U content. However, in accordance with previous observations, analyses of exterior soil samples confirmed high enrichments of ^{226}Ra in karst soils from the Jura Mountains. Von Gunten et al. (1996) attributed these local enrichments to chemical weathering of limestone fragments within the soil column and demonstrated that ^{226}Ra is mainly adsorbed by fractions of humic, amorphous (e.g. ferrihydrites) and oxidic (e.g. goethite) materials. Thus, leaching of the exterior soil cover during precipitation events could lead to major concentrations of the decay product ^{210}Pb in drainage confluences. This is particularly well illustrated at Monlési ice cave by the high activity measured on cryogenic cave calcite (Table 5), a fine carbonate powder issued from the segregation of solutes at slow freezing rates. Adsorption of ^{210}Pb is favoured by the feeding drip-water flowing along the ice speleothems.
Furthermore, organic material present at cave entrances constitutes an ideal trap for ^{210}Pb from atmospheric fallout. This assumption is validated by the analysis of organo-clastic material sampled on the ice surface of both study sites (Table 5). The disequilibrium of the ^{226}Ra/^{210}Pb ratios suggests major enrichments in ^{210}Pb. The presence of ^{40}K and ^{137}Cs (mainly from the 1986 Chernobyl release) confirms that this enrichment comes from atmospheric fallout and is not produced within the cave. This interpretation is validated by high ^{226}Ra activities, suggesting that the samples were in contact with the exterior soil cover. Therefore, the presence of organic material on the ice surface constitutes a potential source of ^{210}Pb enrichment of cave ice.
Moreover, due to the regional enrichment of ^{226}Ra, the

Tab. 5 Radioactivity of clastic material found in Monlési and St-Livres cave ice. The overlying soil is at the origin of most of the radioactivity observed within the caves. (n.d.: not detected)

Sample	U-238 [Bq/kg]
Kimmeridgian limestone (Monlési)	16 ± 9
Reference sample (soil Marchairuz)	63 ± 13
Organo-clastic material (Monlési)	30 ± 18
Organo-clastic material (St-Livres)	< 36
Cryogenic calcite (Monlési)	20 ± 15

decay product ^{222}Rn is present naturally in soils of the Jura Mountains and might be advected by water infiltrations into the karst system (e.g. Surbeck, 1990; Von Gunten et al., 1996). With its short half-life time life (3.82 days), the ^{222}Rn activity of percolating karst water is closely related to the hydrologic history (for instance storage within the epikarst) and hence, constitutes a good natural tracer for hydrological investigations (Eisenlohr & Surbeck, 1995). Representative values of the soil activity are preferentially reached after flood peaks with a time delay proportional to the storage capacity of the aquifer. Typical radon activity within the aqueous phase is in the order of 5 Bql^{-1}, but values up to 50 Bql^{-1} are expected in pore-water of Jura Mountain soils (Surbeck, 1990). Setting the postulate that the cave ice results only from the refreezing of infiltration water, the ^{210}Pb-contamination rate of the ice issued from ^{222}Rn present in water can be assessed at:

$$[^{210}Pb] = \{(t_{1/2}\ ^{222}Rn)/(t_{1/2}\ ^{210}Pb)\} * [^{222}Rn] \quad (1)$$

where $[^{210}Pb]$: lead-210 concentration [Bq/l]; $t_{1/2}\ ^{222}Rn$: half-time life of radon-222; $t_{1/2}\ ^{210}Pb$: half-time life of lead-210; $[^{222}Rn]$: concentration of radon-222 [Bq/l].

From Relation (3) it is shown that for a ^{222}Rn activity of about 10 Bq/l, typical for water inlets in Monlési ice cave (Table 6), the potential contamination represents only a few percent of the ^{210}Pb input from precipitation. But, further contamination could occur via dry ^{210}Pb deposits from exposure to the ^{222}Rn-enriched cave atmosphere. Applying the «volume traps» technique (Oberstedt et al., 1996), Falk et al. (2001) correlated measured surface activity to the exposure time. For a system at equilibrium, these authors suggest an empirical relation between the ^{222}Rn activity in the air and ^{210}Pb deposits:

$$A\ ^{222}Rn / A\ ^{210}Pb = 42\ Bqm^{-3} / 1\ Bqm^{-2} \quad (2)$$

Where $A\ ^{222}Rn$: activity of radon-222 [Bqm^{-3}]; $A\ ^{210}Pb$: activity of lead-210 [Bqm^{-2}]

Observations in Monlési ice cave (table 6) indicate a ^{222}Rn activity in cave air of about 500 $Bqm^{-3.}$ The hypothesis is set that lead deposits originating from the cave air are possible only about 7 months a year (i.e. ~60 % of time), because of melting processes during the remaining time. From Relation (2) the mean annual ^{210}Pb enrichment of the ice is assessed at about 7 Bqm^{-2} in Monlési ice cave. This value is almost insignificant with regard to the natural ^{210}Pb activity of the precipitation water (~170 Bqm^{-2} after Caillet, 1999). However, if ice accumulation rates are low (for instance ≤100 $kgm^{-2}year^{-1}$) this source of contamination cannot be neglected. The same process for exposure to ^{222}Rn-enriched water circulating at the ice surface should be considered.
The large scattering of our analyses led us to infer that ^{210}Pb activity could be closely related to the presence of sediments within the cave ice.

Ra-226 [Bq/kg]	**Pb-210** [Bq/kg]	**K-40** [Bq/kg]	**Cs-137** [Bq/kg]	**Remarks**
20 ± 33	26 ± 7	n.d.	n.d.	The U-series seems to be in equilibrium. The Ra-226/Pb-210 ratio suggests an absence of radon emanations.
343 ± 3	191 ± 15	n.d.	n.d.	The disequilibrium observed between Ra-226 and Pb-210 suggests a strong emanation of radon from these soils.
101 ± 43	386 ± 28	157 ± 30	48.5 ± 4	The imbalance observed between $^{226}Ra/^{210}Pb$ suggests that both samples are «contaminated» in ^{210}Pb. The presence of ^{40}K and ^{137}Cs suggests that this contamination issues from exterior aerosols and is not produced in situ. This interpretation is validated by the high ^{226}Ra activity of the organic material, suggesting that it was in contact with the epikarst.
59 ± 43	1280 ± 47	109 ± 28	211 ± 7	The high concentration of Pb-210 observed in a sample issued from the top of the ice filling suggests that Pb-210 dating methods cannot be applied in this cave. The disequilibrium observed with Ra-226 suggests that this activity results from a long-term exposition to a cave atmosphere enriched in radon.
< 40	641 ± 28	89 ± 22	21.8 ± 2	

Tab. 6a Radon-222 activity of water samples from Monlési and St-Livres ice caves.

Cave	Sample date and time	Station	Water discharge [$lmin^{-1}$]	^{222}Rn (Water) [Bq/l @ sampling]
Monlési	040315 14:20	drip water	<0.2	5.9 +/- 0.7
Monlési	040315 14:15	drip water	<0.2	3.6 +/- 0.6
Monlési	040315 13:50	drip water	<0.2	3.7 +/- 0.6
Monlési	040430 13:50	main water inlet	1	0.3 +/- 0.2
Monlési	040430 15:10	main water inlet	1	0.8 +/- 0.2
Monlési	040601 20:25	main water inlet	1.9	5.4 +/- 0.5
Monlési	040601 21:20	main water inlet	2.5	8 +/- 0.6
Monlési	040601 21:30	main water inlet	5.6	6.2 +/- 0.5
Monlési	040601 21:50	main water inlet	9.1	7.7 +/- 0.6
Monlési	040912 07:50	main water inlet	3.06	11.2 +/- 0.4
Monlési	040912 08:10	main water inlet	2.66	10.7 +/- 0.4
Monlési	040912 08:20	main water inlet	2.60	11.6 +/- 0.4
Monlési	040912 08:30	main water inlet	2.56	10.6 +/- 0.4
Monlési	040912 08:40	main water inlet	2.48	10.3 +/- 0.4
Monlési	040912 08:50	main water inlet	2.37	11.1 +/- 0.4
Monlési	040912 09:00	main water inlet	2.21	12.4 +/- 0.4
Monlési	040912 09:10	main water inlet	1.99	11.3 +/- 0.4
Monlési	040912 09:20	main water inlet	1.84	10.6 +/- 0.4
Monlési	040912 09:30	main water inlet	1.69	10.8 +/- 0.4
Monlési	040912 10:10	main water inlet	1.20	10.4 +/- 0.4
Monlési	040912 10:20	main water inlet	1.14	10.5 +/- 0.4
Monlési	040912 10:30	main water inlet	0.96	10.3 +/- 0.4
Monlési	040912 10:40	main water inlet	0.90	11.2 +/- 0.4
Monlési	040912 10:50	main water inlet	0.83	10.7 +/- 0.4
Monlési	040912 11:00	main water inlet	0.77	9.2 +/- 0.4
Monlési	040912 11:10	main water inlet	0.71	9.6 +/- 0.4
St-Livres	040924 12:10	main water inlet	~10	<1
St-Livres	041021 10:05	main water inlet	21	1.3 +/- 0.3
St-Livres	041021 10:15	main water inlet	0.125	1 +/- 0.3
St-Livres	041021 10:22	drip water	<0.5	1.5 +/- 0.3
St-Livres	041021 10:25	drip water	<0.5	0.6 +/- 0.3
St-Livres	041021 10:30	lateral flowstone	<0.5	1.1 +/- 0.3
St-Livres	041021 10:35	main water inlet	21	0.4 +/- 0.3

Tab. 6b Radon-222 activity of air samples from Monlési and St-Livres ice caves.

Cave	Sample date and time	Station	Water discharge [$lmin^{-1}$]	^{222}Rn (Air) [Bq/m^3 @ sampling]
Monlési	040601 21:20	main room	2.5	520 +/- 33
Monlési	040601 21:50	main room	9.1	488 +/- 31
Monlési	040614 11:50	main room	<1	452 +/- 34
St-Livres	040910 16:30	main room	<1	3000 +/- 90
St-Livres	041021 10:00	main room	21	2600 +/- 70
St-Livres	041021 10:40	main room	21	2760 +/- 70

Our results supported observations made by Von Gunten et al. (1996) and suggest that karst soils may contain high concentrations of ^{226}Ra related to the chemical weathering of limestone. Therefore, due to possible contamination of cave ice by ^{226}Ra, the ^{210}Pb method cannot be applied for the dating of samples with high contents of clastic sediments. Nevertheless, analyses from the Monlési study site demonstrated that ^{210}Pb content of recent clear massive congelation ice (82.2 m Bqkg^{-1}) was consistent with values expected from precipitation water. By considering a slightly enlarged uncertainty related to infiltration water enriched in ^{222}Rn, lower cave ice samples from Monlési ice cave (^{210}Pb: ~15 m Bqkg^{-1}) suggest an age range of between 60 and 90 years, consistent with other observations. Therefore, ^{210}Pb can be considered as a reliable dating method of clear massive congelation cave ice.

3.5 Dating with seasonal fluctuating parameters ($\delta^{18}O$)

Temperature- and vapour-pressure-dependent fractionation of the main isotopic components of water $H_2{}^{16}O$ and $H_2{}^{18}O$ causes lower $\delta^{18}O$ values in winter and higher values in summer precipitations (Dansgaard et al., 1964). Glaciological investigations of high-alpine ice cores demonstrated that these fluctuations might be well-preserved in perennial firn accumulations. Thus, $\delta^{18}O$ signatures enable the distinction of successive annual ice layers and their precise counting provides an excellent approach for the dating of ice cores (e.g. Eichler et al., 2000).
However, the presence of major air circulations could lead to significant fractionation processes in subsurface environments. Luetscher & Jeannin (in press) concluded that evaporation/sublimation is maximal during cold winter days, when the temperature difference between the cave air and the external atmosphere is great. Hence, during the dry cold winter season (i.e. top of annual ice layers), $\delta^{18}O$ values in the ice are significantly increased. Conversely, condensation is chiefly observed during the summer season when humid air flows over a cold substratum.
Oxygen isotope analyses were performed on a reduced section of the ice core extracted from Monlési ice cave. Fifty-eight samples from between 1 m and 1.7 m below the ice surface were analysed. Results show a $\delta^{18}O$ value varying between -7.3 and -12.3 ‰ (Fig. 5) which is consistent with regional precipitation data provided by the NISOT station of La Brévine (Schürch et al., 2003). Although the $\delta^{18}O$ value in precipitation is nearly constant, the oscillating signal observed between -1.3 m and -1.7 m is interpreted as a seasonal signature of the cave air dynamics. Ice crystallization occurs more readily during the late spring when the cave mostly becomes a closed system due to a thermal trap behaviour. This period favours an enrichment in ^{16}O emanating from condensation processes. This interpretation concurs with similar observations achieved in the Canadian Rocky Mountains (Yonge & MacDonald 1999) and the hypothesis is set that low $\delta^{18}O$ values correspond preferentially to early annual ice accumulations. Data from Monlési ice cave suggests that the thickness of annual ice layers ranges between 3 and 10 cm, which corresponds to estimations provided by other methods.

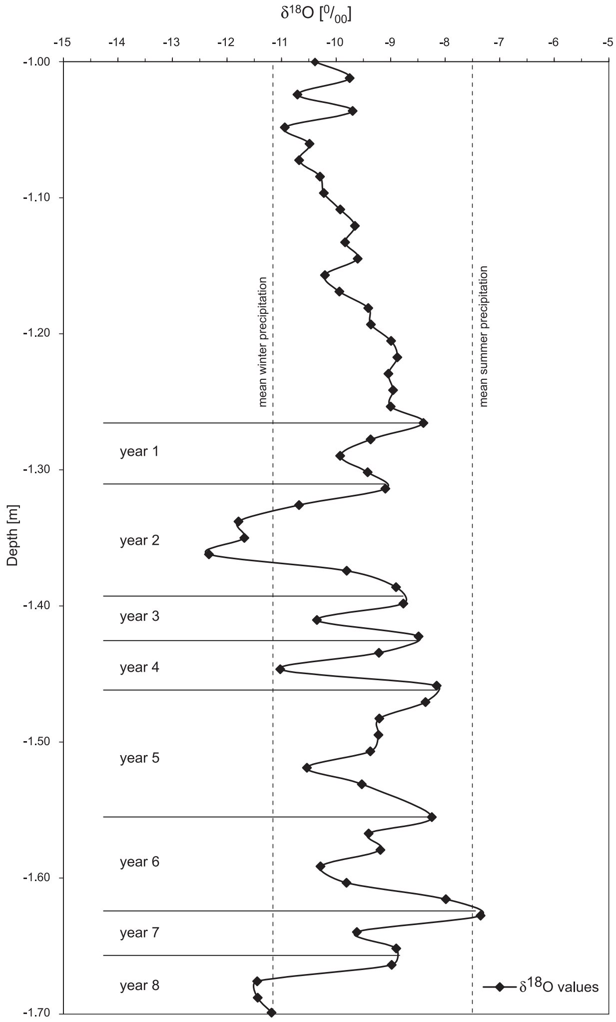


Fig. 5 $\delta^{18}O$ values in ice samples issued from Monlesi ice cave at a depth ranging between -1 m and -1.7 m. Oscillations are attributed to fractionation processes related to cave air circulation. Interpreted ice layers range between 3 and 10 cm, which corresponds to estimations provided by other methods.

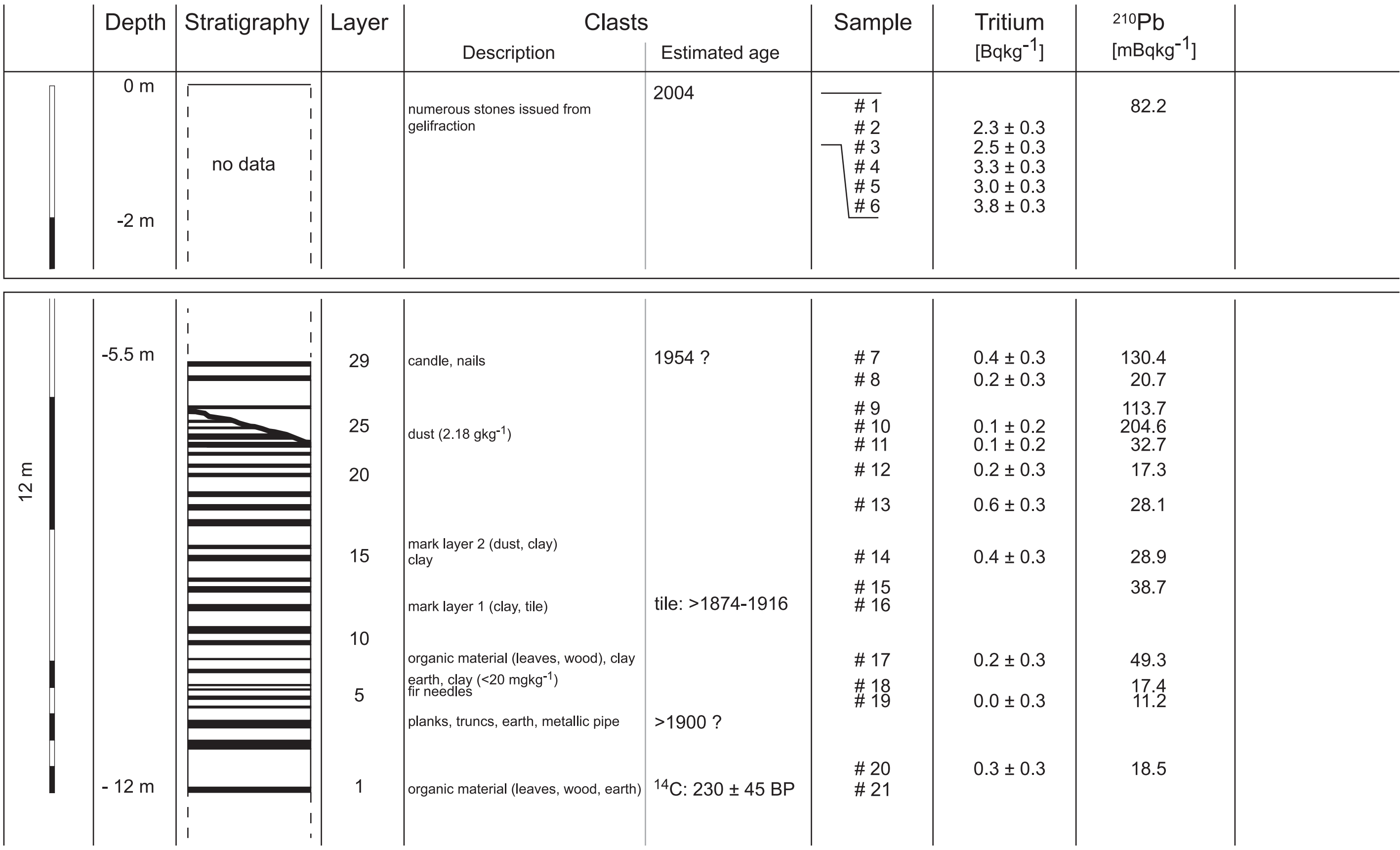


Fig. 6 Synthesis of glaciological investigations achieved in Monlési ice cave. A multi-parametric dating-approach suggests a maximal age of ~120 years for the lowest ice layers.

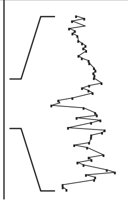
4 | Conclusion

The accessible time period of mid-latitude/low-altitude cave ice is little known. A case study conducted concurrently on two study sites in the Jura Mountains attempted to investigate the accuracy and limitations of several dating methods applied to cave ice. Our results suggest that accurate dating is possible if based on a multi-parametric approach involving isotopic tracers and stratigraphic markers. It was shown that a chronology could be established for recent deposits even in the presence of discontinuous sedimentary profiles.

Figure 6 synthesizes the results compiled from Monlési ice cave, a study site in the Swiss Jura Mountains. Results show that the investigated cave ice has a maximal age of about 120 years . This result implies a fast mass turnover rate induced by a ground heat flux of nearly 1 Wm^{-2} issued from the underlying karst system.

Although recent investigations considered superficial contamination as a critical factor for the reliability of tritium analyses in cave ice (e.g. Pavuza & Mais, 1999; Pavuza & Spötl, 1999), our data demonstrates that this isotope could be used successfully for the identification of young congelation ice deposits (<50 years). However, since tritium enrichments related to regional manufacturing should be considered in low-altitude industrialized environments, more accurate dating using the radiogenic decay of tritium is possible only by calibrating measurements on a local reference curve issued from isotope measurements in precipitation. Comparison data collected since 1994 is provided for the Monlési study site by the NISOT-station of La Brévine (Schürch et al., 2003). Nevertheless, our data confirm that tritium may still represent a good natural tracer for the dating of mark-layers in massive congelation cave ice occurrences (for instance, the 1963 ^{3}H-peak). Similarly, the Chernobyl accident resulted in the fallout of about 64 PBq of ^{137}Cs over Europe and adjacent regions (De Cort et al., 1998). The presence of this isotope on organo-clastic material found within cave ice is therefore a valuable indicator for the identification of cave sediments deposited after 1986.

The analyses of ^{210}Pb represent a valuable qualitative method for the identification of cave ice deposits of less than 200 years. However, due to local enrichments of karst soils in the mother element ^{226}Ra (see also Von Gunten et al., 1996), this method is assumed to be applicable to cave ice only in the absence of exogenous soil material.

$\delta^{18}O$ [$^0/_{00}$] -15 -12 -9 -6	Age [year AD]	Interpretation
	2003 ~1985	**Stable isotopes**: the oscillating $\delta^{18}O$ record is attributed to fractionation processes induced by winter air circulations. Hence, stable isotopes in cave ice can be considered as good proxies for winter temperature fluctuations.
	~1950 ?	**Tritium**: the very low tritium content of ice samples issued from the lower part of the stratigraphy suggests an age of over 50 years.
		Lead-210: the presence of ^{210}Pb in the lower layers suggests an age of less than 250 years. Owing to the contamination induced by radon enriched water circulations, input activity is not controlled. However, an age ranging between 80-120 years can be assessed.
	~1885 ?	**Clasts**: a calibtrated ^{14}C analyse suggests a maximal age of 500 years. The presence of anthropogenic material confirms that cave ice could even be less than about 130 years old.

Furthermore, since ^{210}Pb is almost completely adsorbed onto clay minerals or organic matter (e.g. Wang et al. 1995), accurate dating based on the radiogenic decay of ^{210}Pb is possible only on clear massive congelation ice samples. But even then, it was demonstrated that contamination due to infiltration water enriched in ^{222}Rn must be considered for the final interpretation.

The reliability of radiogenic isotopes for the dating of recent cave ice deposits is validated by the identification of various stratigraphic markers (for instance anthropogenic material). Dendrochronology is a valuable method if material is available in sufficient amount and if calibration is possible on a local reference curve. Despite some uncertainties related to the shape of the calibration curve, radiocarbon is still considered as the most suitable method for a rough approximation of thousand-year-old cave ice deposits. Flow measurements constitute an efficient way to assess mass turnover rates because a steady ground heat flux occurs at the base of the ice filling. Using accurate measurements (<1cmyear^{-1}), estimation of the age of the ice is possible up to several hundred years.

Future investigations should focus on the dating of comparison sites to acquire a general overview of the time-space distribution of the potential archives. A special emphasis should be on the extraction of continuous ice cores which facilitate a high-resolution dating of individual ice layers.

5 | Acknowledgements

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The role of winter air circulations for the presence of subsurface ice accumulations: an example from Monlési ice cave (Switzerland)

Article accepted by *Theoretical and Applied Karstology*

Le rôle des circulations d'air hivernales pour la présence de glace souterraine: un exemple de la Glacière de Monlési.

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Abstract

Investigations carried out in Monlési ice cave document the presence of significant air circulation between the two main cave entrances during the winter season. Air velocities measured in a known cross-section enable assessment of a maximal air flow of more than 10 m^3/s. The resulting annual heat exchange is expressed by the temperature difference between the air inflow and outflow. The quantification of evapo-condensation processes sets the contribution of winter cooling to the final energy balance of the system.

Keywords: ice cave, cave climate, air flow, heat exchange, energy balance, Monlési, Swiss Jura.

Résumé

Les recherches menées en période hivernale dans la Glacière de Monlési mettent en évidence d'importantes circulations d'air entre les deux puits d'entrée principaux. Les vitesses de l'air sont mesurées dans une section connue et permettent d'estimer un débit d'air de plus de 10 m^3/s. L'échange de chaleur annuel est exprimé par la différence de température entre les flux d'air entrant et sortant. La quantification des processus d'évapo-condensation définit la contribution du refroidissement hivernal au bilan énergétique du système.

Mots clés: *glacière, climat souterrain, débit d'air, échange de chaleur, bilan énergétique, Monlési, Jura suisse.*

1 | Introduction

Air circulation was recognised early on as an important factor for heat exchanges between karst systems and outside atmosphere (e.g. Bock, 1913). Subsurface ice patterns probably reflect their most outstanding signature. An investigation of the caves' climatic behaviour is therefore fundamentally important to the understanding of processes at the origin of ice caves. Also, this knowledge contributes to a better assessment of cave ice mass balance and its response to global climate change. In this way, the significance of paleoclimatological studies in cave ice can be better evaluated.
Early studies of alpine ice caves (e.g. Thury, 1861) distinguished two major climatic processes, which were soon qualitatively well described (e.g. Balch, 1900, p.122-126):

1 «cold air trap»:
A cave with a single entrance may act as a thermal trap (Fig. 1a). Due to density differences, exchanges will be restricted to an «open» period that corresponds more or less to the winter season ($T°_{surf} < T°_{cave}$) in a descending conduit. In «summer» ($T°_{surf} > T°_{cave}$), convection cells will be limited to the lower part of the cave and a stratification of the subterranean air will be observed, preserving a low temperature inside the cave ($T°_{cave}$< Mean Annual Air Temperature, MAAT).

2 «chimney effect»:
In underground voids with several entrances, air circulation occasioned by a «chimney effect» can be observed (Fig. 1b). In winter, when the outside temperature is lower than the cave temperature, cold air will be sucked in at the lower entrance of the system (driving pressure, $\Delta P_m > 0$), freezing potential water infiltrations. In summer, when surface temperature becomes warmer than the cave temperature, the process is reversed ($\Delta P_m < 0$), and warm air is sucked in at the upper entrance. Therefore, a thermal anomaly compared to the MAAT can be observed in both entrance zones, that is, a «cold» anomaly at the lower entrance and a «warm» one at the upper entrance.

With the recent theoretical contributions of Badino (1995) and Lismonde (2002), processes at the origin of subsurface air circulation are now physically well described. However, only few field investigations have been conducted with the purpose of quantifying the heat transfers resulting from subsurface air circulation.

The present research aims at conducting new field observations on air circulation in the Glacière de Monlési (Switzerland). Results display initial quantifications of the heat exchange induced by winter air circulation. Next to the validation and improvement of our preliminary conceptual model (Luetscher et al., 2003), this approach should contribute to a better modelling of ice caves' distribution in a spatio-temporal context. Actually, linking the presence of cave ice to outside climatic parameters will provide valuable indicators of climate fluctuations during recent centuries.

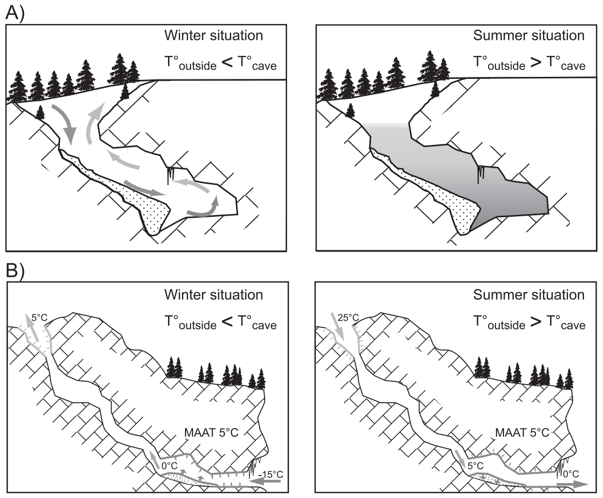

Fig. 1 Schematic behaviour of air circulation in an ice cave. **(A)** Cold air trap. During winter season, when the outside temperature is lower than the cave temperature, the light «warm» air of the cave rises up out of the cave and is replaced by the outside colder air. During summer season, due to density differences, cold air is trapped in the cave and a stratification of the subterranean air is observed. **(B)** Chimney effect. During winter season, when the outside temperature is lower than the cave temperature, the light «warm» air of the cave rises up, out of the cave, and is replaced by cold air sucked into the system at the lower entrance. This process is inverted during summer season. As much more energy is stored in the form of ice (latent heat) than in the rock (sensible heat), the temperature anomaly is larger at the lower entrance.

Fig. 1 *Comportement schématique des circulations d'air dans une glacière.* **(A)** *Piège à air froid. Durant la saison hivernale, lorsque la température extérieure est inférieure à la température de la grotte, l'air chaud de la grotte (moins dense) s'échappe et se trouve remplacé par de l'air frais extérieur. En été, la différence de densité permet le piégeage d'air froid et une stratification de l'air souterrain est observé.* **(B)** *Effet cheminée. Durant la saison hivernale, lorsque la température extérieure est inférieure à la température de la grotte, l'air chaud de la grotte (moins dense) s'échappe et est remplacé par de l'air frais aspiré à l'entrée inférieure. Ce processus est inversé durant la période estivale. L'énergie stockée sous forme de glace (chaleur latente) étant plus important que celle stockée dans la roche (chaleur sensible), l'anomalie de température est plus importante à l'entrée inférieure.*

2 | Site description

Located in the Swiss Jura Mountains (Boveresse/Ne, 6°35'4''/46°56'18'', 1135 m), the «Glacière de Monlési» is presently the largest ice cave known in this mountain range. Early descriptions of Monlési were supplied by Browne (1865). He was the first to observe air oscillations at the cave entrance during the summer season. A detailed cave map by Stettler & Monard (1960) and Stettler (1971) provided new observations on the ice filling.

Three entrance shafts lead to a large room of approximately 25 m x 45 m at a depth of 20 m below the surface (Fig.2). Partially filled with an important mass of congelation ice, the lowest point of the cave can be reached at -33 m through a large «rimaye» located on the south-western part of the «glacier». Luetscher & Wenger (2002) estimated the total ice volume to be about 6000 m^3, but precise measurements of the ice thickness are still lacking.

According to the general nomenclature of ice caves (Luetscher & Jeannin, submitted), Monlési ice cave belongs to the group of statodynamic caves with congelation ice and firn.

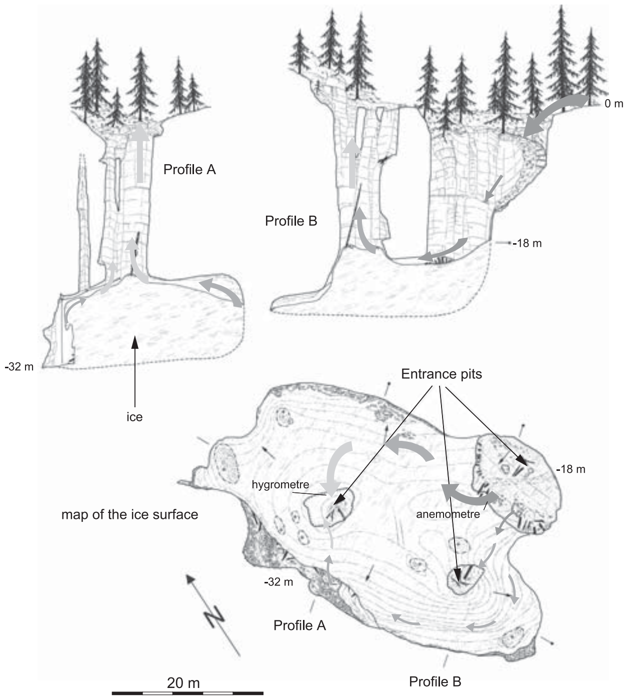

Fig. 2 Plan view and vertical cross-sections of the «Glacière de Monlési» – adapted from Luetscher & Wenger, 2002. Air circulation during the open period is schematised by the arrows. Cold air drops inside the main shaft, warms up in the cave and is blown out from the two other shafts. The position of the minilogger is mentioned by an *.

Fig. 2 *Plan et coupes verticales de la «Glacière de Monlési» – adapté de Luetscher & Wenger, 2002. Les circulations d'air durant la période ouverte sont schématisées par les flèches. L'air froid gravite dans le puits principal, réchauffe la grotte et s'échappe par les deux autres puits. La position du minilogger est mentionnée par une *.*

3 | Methods

In order to characterize the subsurface climatic regime of Monlési ice cave, cave air temperatures were monitored during one annual cycle. Records started on the first of July 2002 and stopped on the 30th June 2003. Continuous logging, at two-hour intervals, was performed with UTL-1 Miniloggers (Hoelzle et al., 1999), previously calibrated in melting ice. Furthermore, continuous logging, at one-hour intervals, was performed with a Campbell CR10X datalogger and two thermistor chains in order to acquire a good overview of the temperature distribution. The 45 «YSI 44006» sensors, spaced at two-metre intervals, were previously calibrated in melting ice. Final accuracy is given at ±0.1 °C. Next to these measurements, field observations were necessary in order to better characterize winter air circulation related to forced convection. Local natural convection cells were neglected as they are assumed to play only a minor role in global heat exchanges. The use of smoke sticks enabled documentation of most of the air flow patterns.
Air velocities were determined with a digital thermo-anemometer (Testo 425) in 22 different points of the cave entrance. Data were then integrated with its cross-section area in order to compute the total air flow. The instrument's accuracy is given at 0.1 m/s and precision of the calculated air flow is estimated at ±10%.
Evapo-condensation processes were investigated with two mirror-type dew point hygrometers (Thygan VTP 37) placed at each extremity of the cave room, within the air flow. Power was supplied for both instruments by 12 V lead accumulators. The recording interval set by the manufacturer is 10 minutes and accuracy is given at ±0.15 °C (i.e. H_r ±2%).

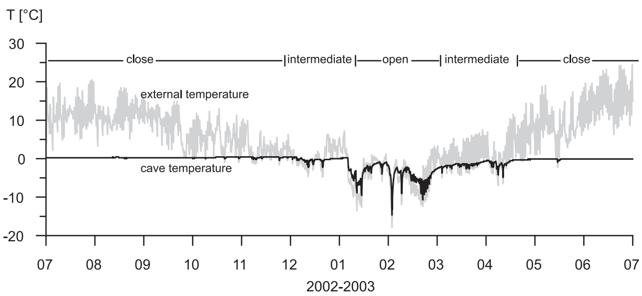

Fig. 3 One year's temperature records acquired at Monlési ice cave. Two periods are outlined in which the cave air temperature displays distinct signatures. During winter season, a good correlation between cave air temperature and negative outside air temperature can be observed. In summer, very stable temperatures, close to 0 °C, are related to the phase changes of ice such as melting.

Fig. 3 *Chronique annuelle des températures de l'air à la Glacière de Monlési. Deux périodes peuvent être observées, au cours desquelles la température de la grotte possède une signature distincte. Durant la saison hivernale, une bonne corrélation est observée entre la température hypogée et la température négative extérieure. Durant l'été, les températures très stables, proches de 0 °C, sont attribuables au changement de phase de la glace (en l'occurrence de la fonte).*

4 | Results and interpretations

4.1 Climatic regime of Monlési ice cave: chimney effect versus cold air trap

Temperatures recorded at Monlési ice cave during the annual cycle 2002-2003 show two periods in which the cave air temperature displays distinct signatures (Fig. 3):
1. The first period, corresponding mostly to the winter season (November to April), is characterized by a good correlation (r^2=0.91) between cave air temperature and negative outside air temperature. When the outside temperature is lower than that of the cave atmosphere, the density difference between the two air masses leads to a chimney effect between the entrances (Fig 2). This forced convection occurred approximately 1150 hours during the 2002-2003 annual cycle. Because of the cave geometry (respective dimensions of the conduits), air flows always in the same direction: it enters inside the cave by the larger shaft and is blown out by the two others.
2. The second period, from May to October, shows a very stable temperature, close to 0 °C. This almost constant value is controlled strictly by the phase changes of ice, in this case related to melting processes. During this period, when outside T° is higher than cave T°, no significant air circulations can be observed: Monlési ice cave acts as a thermal trap (Choppy, 1986). However due to the presence of several entrance shafts, a slight oscillating draft first mentioned by Browne (1865) can be observed. This phenomenon is documented by temperatures recorded at the cave entrance. Lismonde (2001) attributes this characteristic to a pendulum movement related to the presence of several entrances.
According to these observations it can be assumed that, during winter season, air circulation within Monlési ice cave are controlled by a chimney effect. This behaviour enables a major heat exchange with the outside atmosphere. Unlike a usual chimney effect, the process is not inversed during summer season because there is no altitude difference between the different entrances. Thus, the trapping of cold air contributes significantly to the conservation of the cave ice.

4.2 Sensible heat exchanged by forced convection

It has been demonstrated that in forced convection, air flow is proportional to the square root of temperature difference between the cave and the outside air: $\sqrt{\Delta T}$
(e.g. Lismonde, 2002). Actually, velocities measured at the cave entrance show an excellent correlation with modelled values (reg. coeff.: 0.98, Fig. 4). Using continuous temperature records, the total air flow through Monlési ice cave could be computed for the annual cycle 2002-2003. A mean equivalent air flow during the open period (November to April) of about 3 m^3/s was determined.

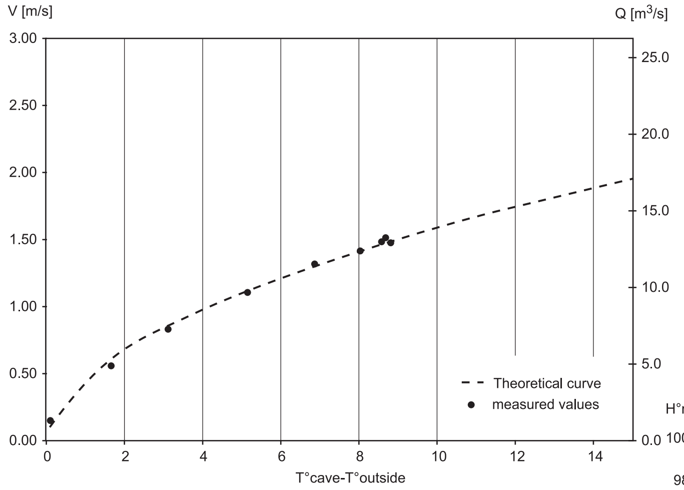

Fig. 4 Air flow measured at the main entrance of Monlési ice cave. Measured values fit well with the theoretical curve.

Fig. 4 *Débit d'air mesuré à l'entrée principale de la Glacière de Monlési. Les valeurs mesurées corrèlent bien avec la courbe théorique.*

Furthermore, the temperature difference between air inflow and outflow was measured at two of the entrance pits (Fig. 5). Therefore, the sensible heat lost by the cave through forced convection is obtained by integrating the following relation:

$$H = \int_0^{time} Q_{air} \cdot c_{air} \cdot \left(T_{out} - T_{in}\right) dt \tag{1}$$

Where: H: lost energy [W]; Q_{air}: specific air flow [kg/s]; c_{air}: heat capacity of air [J/kg]; T_{out}: temperature of air blown out of the cave [°C]; T_{in}: temperature of air sucked into the cave [°C].

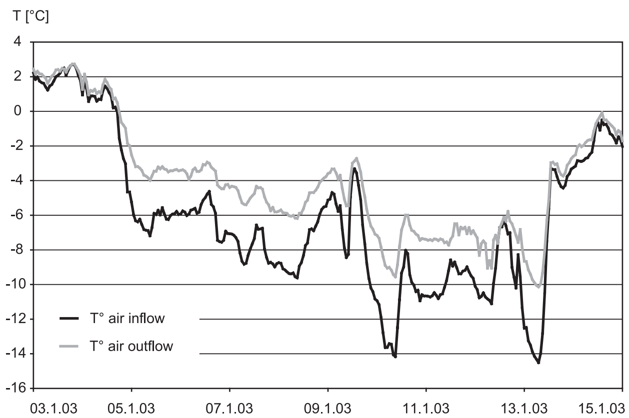

Fig. 5 Air temperatures recorded at two entrance shafts of Monlési ice cave. The heat exchanged by forced convection is obtained by integrating temperature differences between air inflow and outflow.

Fig. 5 *Températures de l'air enregistrées à deux puits d'entrée de la Glacière de Monlési. La chaleur échangée par convection forcée est obtenue par l'intégration des différences de température entre l'air entrant et l'air sortant.*

The total energy exchanged with outside atmosphere during the 2002-2003 cycle is estimated to 12 MWh. Setting the (simplistic) hypothesis that the entire heat loss is attributed to ice crystallisation, it is possible to assess the maximal seasonal ice deposit at about 140 m^3 in 2002-2003. This value is roughly in the same order as observed in the field (~10 cm/m^2). However, it must be considered that a part of the exchanged energy has not been transformed into ice but was stored in the rock and in the ice. Further heat transfers related to evapo-condensation processes must also be assumed.

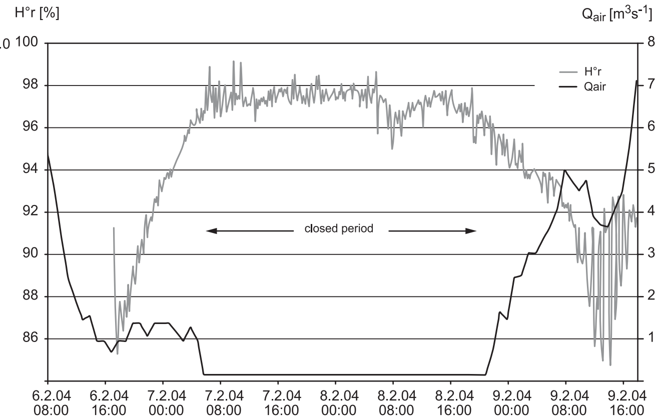

Fig. 6 Air velocities and related cave air humidity inside Monlési ice cave. Ventilation episodes transport non-saturated air into the cave, therefore enabling evaporation processes.

Fig. 6 *Vitesses de l'air et humidité dans la Glacière de Monlési. La convection forcée transporte de l'air non-saturé au sein de la cavité et engendre de l'évaporation.*

4.3 Latent heat exchange due to evapo-condensation processes

Cave atmosphere in Monlési ice cave is almost at saturation during the closed period. However, as illustrated on Figure 6, air circulation considerably reduces the saturation of subsurface atmosphere by the transport of colder (i.e. dryer) outside air through the cave. By the way, evaporation/sublimation processes might occur, contributing to the cooling of the cave. These processes are illustrated in the cave by a small insoluble deposit found on the ice surface, coming from the precipitation of ions dissolved in the water. Conversely, hoar frost deposits can be observed on the cooled substratum just after an open period.
Unfortunately, due to technical limitations, continuous measurements of air humidity could not be assessed over a long time period. It seems reasonable, however, to admit that some local effects of sublimation are compensated for by the condensation that follows. These heat transfers due to phase changes of H_2O need to be investigated further.

5 | Conclusions

Investigations carried out in Monlési ice cave enable a better quantification of heat exchange during the 2002-2003 annual cycle. During this period, the maximal ice accumulation could be estimated at 140 m^3, which is in agreement with field observations. Thus, such a value demonstrates that ice accumulation is still a contemporary phenomenon. However, a negative mass balance could be observed during the last decade. These observations correlate well with an increasing winter temperature (Rebetez, 2002) and lower snow precipitations.

Consequently, the hypothesis is set that under climatic conditions like those in the nineties, Monlési ice cave cannot present an equilibrated energy balance.

This study contributes to a better quantification of the energy balance of the Monlési ice cave. Nevertheless, it is important to emphasize that uncertainties remain with the measured air flow since residual air circulation towards the second entrance pit has not been considered yet. Furthermore, no distinction has been made between heat transfers occurring in the form of latent heat and those occurring in the form of sensible heat.

The approach outlined in this study enables an initial approximation of the heat exchange induced by cold air circulations. Further studies should focus on modelling cave temperature variations with regard to outside meteorological records. Linking latent heat exchanges to this model will constitute the final step in this study and should enable a reconstruction of the mass balance of Monlési ice cave during the last century.

6 | Acknowledgements

The authors would like to thank F. Bourret and all the people who helped in the instrumentation of the study site. Our gratitude is dedicated also to W. Haeberli for his participation and advice in this project. This work is part of a larger study on ice caves in the Jura Mountains supported by the Swiss National Science Foundation, Project 21-63764.00.

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Modelling heat transfers in Monlési ice cave

article in preparation

Marc Luetscher, Pierre-Yves Jeannin, Baudouin Lismonde

Abstract

In order to predict the evolution of ice caves in a changing climate context, physical processes at the origin of mid-latitude/low-altitude cave ice were investigated thoroughly at a selected study-site in the Jura Mountains. Measured heat exchanges were validated by simple numerical models and concluded that fluctuations of the mass balance of cave ice are chiefly related to the winter cooling induced by air circulation and the latent heat supplied by intrusive snow accumulations.

Key words: karst, modelling, heat transfer, ice cave, permafrost.

1 | Introduction

Although the presence of cave ice has been documented worldwide, process-based descriptions relying on quantified energy balances remain rare. Subsequently, only few scientific interpretations enable one to predict the evolution of cave ice mass balances in a changing climate context. While an increasing number of studies are concerned with potential paleoclimatic records, the physical processes at the origin of cave ice are usually insufficiently described, strongly limiting the understanding and the interpretation of data. Our study aims at providing a quantification of heat exchanges observed in the mid-latitude/low-altitude Monlési ice cave (1130 m a.s.l., Swiss Jura Mountains). The main issues lie in a precise identification of the meteorological parameters controlling the cave ice mass balance.

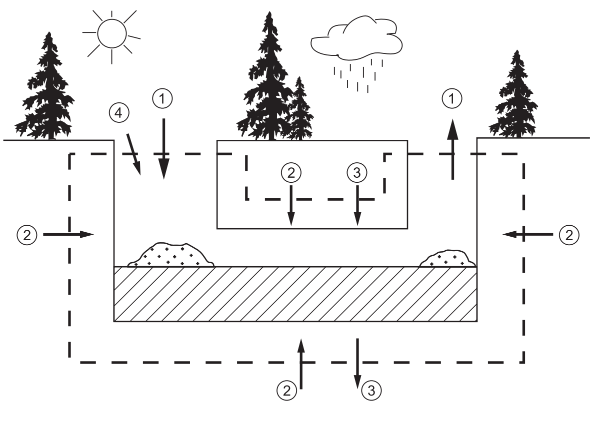

Fig. 1 General system of an ice cave. Heat exchanges observed at system boundaries result from (1) forced convection, (2) conduction through the surrounding limestone, (3) advection by water circulations , and (4) solar radiation.

2 | Methods

The investigated system is comprised of the entire «thermal anomaly» related to the ice cave (Fig. 1). Therefore, it includes the ice, the water passing through the cave (snow, water and vapour), the cave air, and the limestone surrounding the cave up to a distance where the rock temperature remains stable during the entire year. Boundary conditions are defined by **a)** a constant rock temperature, and **b)** exterior meteorological fluctuations above the entrance shafts (daily and seasonal oscillations). Heat exchanges at the boundaries are distinguished by:

1 forced convection;
2 conduction through the surrounding limestone;
3 advection by water circulations;
4 thermal radiation.

The thermodynamical referential is taken at the energy equivalent of melting pure ice. This supposes a zero enthalpy for dry air at 0 °C, a zero enthalpy for water at 0 °C, a positive enthalpy for water vapour and a negative enthalpy for ice and snow (latent heat). The annual energy balance of the global system can be expressed as:

$$\Delta E_{air} + LE_{snow} + \Delta LE_{ice} = S + \Delta E_{water} + R \quad (1)$$

Where ΔE_{air}: heat advected by air circulation; LE_{snow}: latent heat of intrusive snow; ΔLE_{ice}: latent heat of ice; S: ground heat flux; ΔE_{water}: heat advected by water circulation; R= solar radiation.

ΔLE_{ice} represents the adjusting parameter related to the cave ice mass balance. As many of these heat fluxes are difficult to assess, the global system was subdivided into several smaller sub-systems (air, ice and rock), which could be validated more easily by field measurements. These sub-systems are composed of more or less constant volumes. Since they can be passed through by air and/or water fluxes they have to be considered as open sub-systems.

3 | The sub-system air

3.1 Introduction

Field observations performed in Monlési ice cave confirm the presence of a significant unidirectional air circulation on cold winter days. Conversely, during the summer season this ventilation is reduced to a small oscillating air flow. These observations suggest the presence of two distinct phases, the one corresponding to an open system while the other looks more like a «closed» system. Figure 2 illustrates air temperature measurements acquired during a three-year monitoring at Monlési ice cave. The recorded data highlights a period ranging from November to April which is characterized by a good correlation (r^2=0.91) between the cave air temperature and the negative exterior air temperature. This typical signature suggests that air circulation is closely related to density differences between the cave air and the exterior atmosphere.

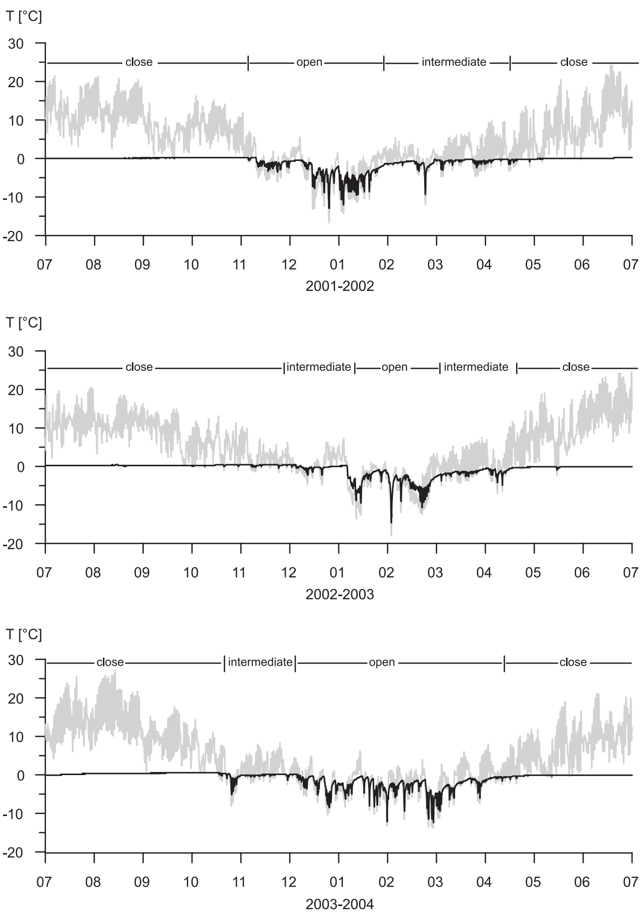

Fig. 2 Air temperature recorded at Monlési study-site between 07.2001 and 07.2004. Recorded data highlight an open period ranging between November and April which is characterized by a good correlation between the cave air temperature (black) and the negative exterior air temperature (grey).

Luetscher & Jeannin (in press) estimate that the system was open for approximately 1150 hours during the 2002-2003 annual cycle. In order to quantify the heat transfer associated with this air circulation, the heat balance of the sub-system air was the object of a detailed investigation, which is presented below.

3.2 Definition of the sub-system

The sub-system air is defined by considering only the cave atmosphere, of more or less constant volume. As it can be crossed by air and water flows it is considered to be an open sub-system. Heat transfers at boundaries include solar-, rock- and ice- radiation, phase changes issued from evapo-condensation processes (sublimation and condensation), advected energy related to air flows, and advected energy related to water flows. Since radiation affects the rock and ice surfaces rather than the air itself, it is assumed that this term can be neglected in the energy balance of this sub-system. Hence, the annual energy balance is given by:

$$LE_{cond.} + E_{air} + A_{water} = LE_{sub.} + E'_{air} + A'_{water} + \Delta H \qquad (2)$$

Where $LE_{cond.}$: latent heat related to condensation; E_{air}: heat advected by air into the system; A_{water}: heat advected by water into the system; LE_{sub}: latent heat related to sublimation; $E'_{air:}$ heat advected by air out of the system; A'_{water}: heat advected by water out of the system; ΔH: energy lost by the sub-system in contact of the surrounding rock and ice.

It must be noted that $LE_{cond.}$ and LE_{sub} do not represent all phase changes present in the ice cave since those are mostly local processes. In fact, these latent heat transfers are distributed on both bodies considered (for instance air/rock, air/ice). Their quantification would require the study of the local thermics.
The following developments aim to provide a field-based quantification of each term intervening in the energy balance of the sub-system (2). Air and water flows are reconstructed in order to assess the heat exchanged during an annual cycle. Fluctuations of the internal energy of the sub-system are interpreted as closely related to the mass balance of the ice cave.

3.3 Air dynamics

3.3.1 Laminar versus turbulent flow

A fluid moves either in laminar or turbulent flow conditions depending on the roughness of the boundary layer and the fluid's velocity. Conversely to laminar flows, fluid motion in a turbulent medium is highly irregular and is characterized by velocity fluctuations. These fluctuations enhance the transfer of momentum, energy and species, and hence increase surface friction as well as convection transfer rates.

Transition between laminar and turbulent flow conditions is characterized by the Reynolds number (Re). In circular conduits, this number is expressed as:

$$\mathrm{Re} \equiv \frac{VD}{\nu} \tag{3}$$

Where Re: Reynolds number; V: mean fluid velocity [m/s]; D: aeraulic diameter (D=4S/P) [m]; S: cross section area [m^2]; P: section perimeter [m]; ν: kinematic viscosity [m^2s^{-1}].

The kinematic viscosity of air can be computed using the expression given by Netz (1983):

$$\nu = \frac{1}{p} \cdot \frac{4.87 \cdot T^{1.5}}{1+\frac{172.6}{T}} \cdot 10^{-6} \tag{4}$$

Where ν: kinematic viscosity of air [m^2s^{-1}]; p: air pressure [hPa]; T: temperature [K]. (air at 0 °C about $14*10^{-6}$ m^2s^{-1}).

In smooth industrial pipes, flow is fully laminar up to a critical Reynolds number of Re ≈ 2300 (e.g. Incropera & De Witt, 2002). Lismonde (2002, p. 53) suggested that this value might be significantly lower for natural karst conduits (i.e. ~640). In this context (i.e. ε/D > 0.05, where ε is the roughness of the conduit and D the aeraulic diameter), fully turbulent flow conditions are achieved when Re > 10000 (Lomize, 1947; Louis, 1968; Nékrassov, 1968 in Jeannin & Maréchal, 1995). For 640 < Re < 10000, a transition between laminar and turbulent flow conditions is observed.

During open system conditions, air flow measurements performed in Monlési ice cave provided values ranging between 1 and 15 m^3s^{-1} (Luetscher & Jeannin, in press). In the largest parts of the cave (D ≈ 5 m), air velocities range between 0.05 and 0.8 ms^{-1}. Subsequently, the Reynolds number ranges between $1.8 \cdot 10^4$ and $2.8 \cdot 10^5$, which implies a fully turbulent air flow. According to the Moody diagram (Moody, 1944) the flow within Monlési ice cave is in the fully rough zone if Re > 45'000, which is assumed when the air flow is higher than 2.5 m^3s^{-1}.

3.3.2 Forced convection

As illustrated by Figure 3, the cold exterior air reaches the cave by the main entrance shaft, warms up in contact with the rock and the ice, and is blown out from the two other shafts. This forced convection is attributed to the density difference between the subterranean air and the exterior atmosphere. Since air moisture plays only a secondary role compared to the temperature difference between the cave and the exterior, and considering an absence of significant CO_2 concentrations in the cave atmosphere, a physical approximation of the driving pressure is provided by Equation (5) (Lismonde, 2002):

$$\Delta P_m \approx \rho_0 \frac{p}{p_0} \frac{g}{273} (T_{\mathrm{int}} - T_{ext}) H \tag{5}$$

Where ΔP_m: driving pressure [Pa]; ρ_0: mean density of air [kg/m^3]; p: cave air pressure [Pa]; p_0: air pressure at surface [Pa]; g: earth acceleration [m/s^2]; T_{int}: mean cave air temperature [K]; T_{ext}: mean exterior temperature [K]; H: altitude difference [m].

The air flow given by the turbulent flow equation (Darcy-Weisbach):

$$q_m = \sqrt{\frac{|\Delta P_m|}{R}} \tag{6}$$

where q_m: air flow through the system [kg/s]; ΔP_m: driving pressure [Pa]; R: aeraulic resistance of the conduit [$kg^{-1}m^{-1}$].

The aeraulic resistance (R) reflects the headlosses occurring in a pipe. Jeannin (2001) suggested that in common karst conduits regular headlosses usually dominate over singular ones. Thus, in a fully developed turbulent flow, the aeraulic resistance is given by (e.g. Lismonde, 2002, p. 57):

$$R = \frac{8 \Lambda L}{\pi^2 \rho D^5} \tag{7}$$

Where R: aeraulic resistance of the conduit [$kg^{-1}m^{-1}$]; Λ: headloss coefficient; L: conduit length [m]; ρ: density of air [kgm^{-3}]; D: aeraulic diameter (D=4S/P) [m].

The dimensionless headloss coefficient Λ is a function of the Reynolds number for transition flows (640 < Re < 10000) but depends only on the relative roughness ε/D of the conduit in fully turbulent flows. Jeannin & Maréchal (1995) considered Louis' empirical relation (Louis, 1968) as the most adapted for the determination of headloss coefficients in regular karst conduits:

$$\frac{1}{\sqrt{\Lambda}} = -2 \log\left(\frac{\varepsilon}{1.9D}\right) \tag{8}$$

Where Λ: headloss coefficient; ε: roughness of the conduit [m]; D: aeraulic diameter (D=4S/P) [m].

Observations carried out in Monlési ice cave outline a bedrock roughness mostly controlled by the lithology and the limestone dissolution. Intense gelifraction leads to numerous stone deposits on the top of the ice filling, which might also be the object of tension cracks or small scale «bédières». Thus, considering a mean roughness of 50 cm appears to be reasonable. Applying equation (8) to the Monlési ice cave enables assessment of a headloss coefficient Λ of ~0.2, consistent with values suggested by Lismonde (2002, p. 54).

Air flow measurements taken in Monlési ice cave confirm the linear correlation (r^2: 0.99; 9 values) between measured air velocities and the square root of the temperature difference betwee the exterior atmosphere and the cave air (Fig. 4). Consequently, an aeraulic resistance of 0.04 $kg^{-1}m^{-1}$ can be considered as reliable.

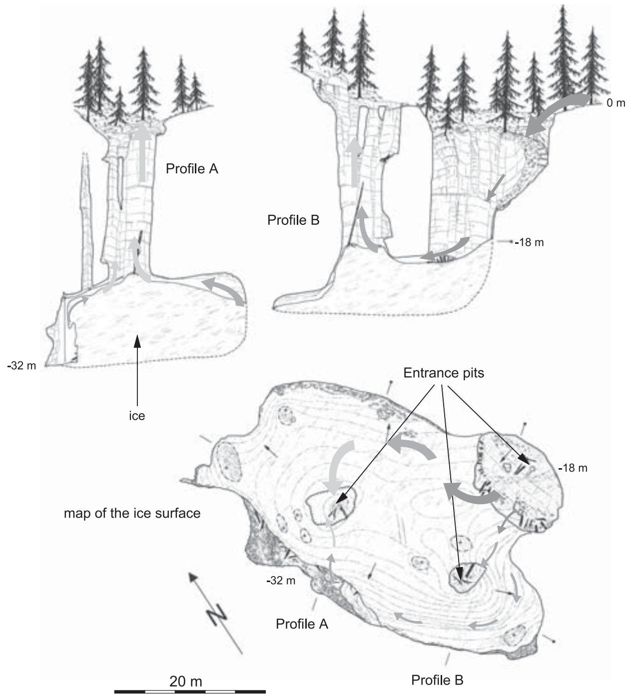

Fig. 3 Forced convection observed in Monlési ice cave during an open period. The cold exterior air reaches the cave by the main entrance shaft, warms up in contact with the rock and the ice, and is blown out from the two other shafts (cave map adapted from Luetscher & Wenger, 2002).

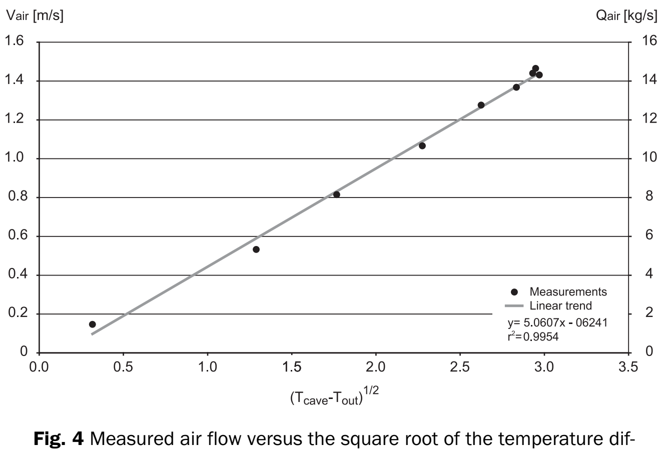

Fig. 4 Measured air flow versus the square root of the temperature difference between cave air and the exterior atmosphere.

3.3.3 Reconstructing the annual air flow

Applying equation (6) to the temperature recorded at Monlési ice cave enabled the reconstruction of the air flow. The computed air volume flowing through the cave during the annual cycle 2002-2003 reached nearly $14*10^6 m^3$, which supposes a mean air flow of about 3.3 m^3s^{-1} (Luetscher & Jeannin, accepted b; Part II, p. 93). This value is consistent with field observations achieved during the open period.

3.4 Heat transfer in turbulent pipes

Setting the hypothesis of a driving pressure (i.e. air velocity) which is dependent only on the exterior air temperature, it is possible to uncouple the thermal aspect from the problem of motion.

In a system defined by air circulating through a pipe, and neglecting the temporal derivative of the internal energy of the air, incoming energy fluxes equal the outgoing fluxes:

$$P_{in} + \Phi_{tot.} = P_{out} \qquad (9)$$

Where P_{in}: power of inflowing air [W]; $\Phi_{tot.}$: heat flux supplied to the system [W]; P_{out}: power of outflowing air [W].

3.4.1 Determining the power of inflowing air:

Considering humid air as a mixture of ideal gases, the specific enthalpy h_s is given by:

$$h_s = \frac{H}{m_{as}} = (c_{p_a} + \omega c_{p_v})\,\theta + \omega L \qquad (10)$$

$$h_s = (1.006 + 1.826\,\omega)\,\theta + 2.5 * 10^3\,\omega$$

and,

$$P_{air} = Q_{air}\, h_s \qquad (11)$$

Where *H*: enthalpy of air [kJ]; m_{as}: mass of dry air; h_s: specific enthalpy of humid air [kJ/kg]; c_{pa}: heat capacity of dry air at constant pressure [$kJkg^{-1}K^{-1}$]; c_{pv}: heat capacity of vapor at constant pressure [$KJkg^{-1}K^{-1}$]; ω: specific humidity [kg/kg_{as}]; θ: temperature [°C]; P_{air}: power of circulating air [W]; Q_{air}: massic air flow [kgs^{-1}].

According to humidity records from four MeteoSwiss weather stations, the hypothesis was set that the exterior relative air humidity (H°r) at Monlési site is more or less equal to that measured in the nearby MeteoSwiss station of La Chaux-de-Fonds. This assumption was validated by occasional field measurements. Thus it is possible to compute the specific enthalpy of the air flowing inside the cave. Data from the 2002 - 2003 annual cycle provide an order of magnitude of - 60 GJ transferred towards the system by air convection during the open (winter) season. Although it was shown that during the summer season ventilation processes are almost insignificant, a small oscillating air flow is observed between the different entrance shafts (i.e. ~1 kgs^{-1}). Since measurements concluded that this air flow induced a temperature fluctuation of about 0.2 °C, the energy supplied to the sub-system during the «closed» season was assessed to be approximately + 6 GJ.

Major uncertainties are related to the determination of the specific humidity and the computed air flow. Hence, final accuracy of the computed energy supply was estimated at ± 30 %.

3.4.2 Convective heat exchange along a conduit:
The energy exchanged along a pipe ($\Phi_{tot.}$) can be expressed as the sum of the sensible heat flux issued from the pipe wall and the latent heat flux related to evapo-condensation processes:

$$\Phi_{tot.} = \Phi_{sens.} + \Phi_{lat.} \tag{12}$$

Heat transferred by conductive exchanges at walls ($\Phi_{sens.}$) is given by Newton's law of cooling, where:

$$\varphi = h_{th} (T_1 - T_2) \tag{13}$$

and,

$$h_{th} = (\lambda / D)\ Nu \tag{14}$$

With φ: density of heat flux [Wm^{-2}]; h_{th}: convection heat transfer coefficient [$Wm^{-2}K$]; λ: heat conduction of air [$Wm^{-1}K^{-1}$]; D: aeraulic diameter (D=4S/P) [m]; Nu: Nusselt number; T_1: temperature of the wall; T_2: mixed air temperature.

In a thermally fully developed flow with constant properties, the local convection coefficient h_{th} is constant, independent of the length of the conduit (e.g. Incropera & DeWitt 2002). Empirical data (e.g. Colburn 1933) have shown that Nusselt number can be approximated by:

$$Nu = \alpha_{conv.} Re^{0.8} Pr^{1/3} \tag{15}$$

where

$$Pr = \nu / a \tag{16}$$

With $\alpha_{conv.} \approx 0.046$ for a roughness close to natural karst conduits (Florsch 1995 in: Lismonde 2002, p. 78); Re: Reynolds number (VD/ν); Pr: Prandtl number (0.71 for air); ν: kinematic viscosity of air [m^2s^{-1}]; a: thermal diffusivity of air [m^2s^{-1}].

Hence, the heat flux transferred in a circular conduit is given by (Lismonde 2002, p.77):

$$\Phi_{sens} = 3.57\, \alpha_{conv}\, \lambda \left(\frac{q_v}{\upsilon} \right)^{0.8} \frac{S^{0.2}}{D^{1.2}}\, L \Delta T \tag{17}$$

Where Φ: heat flux [W]; α_{conv}: convection coefficient; λ: thermal conductivity of air [$Wm^{-1}K^{-1}$]; q_v: air flow [m^3s^{-1}]; ν: kinematic viscosity of air [m^2s^{-1}]; D: aeraulic diameter (D=4S/P) [m]; L: length of the conduit [m]; ΔT: temperature difference [K].

In order to compute the energy balance of the air, two distinct models were considered. The first, disregards latent heat exchanges by considering that the energy required for the phase change is provided by a constant wall temperature. In the second model, latent heat exchanges are considered by setting the hypothesis of a saturated outflowing air.

Model 1
(Lismonde 2002, p. 168; Incropera & DeWitt 2002, p. 477; Wigley & Brown 1971)
By disregarding latent heat exchanges (i.e. enthalpy variations are attributed to temperature variations only), the temperature distribution along a conduit can be determined by (e.g. Lismonde 2002):

$$\frac{\partial T}{\partial x} = 4 \frac{h_{th}}{\rho c_p DU} \left(T_{in} - T_{out} \right) \tag{18}$$

After integration, this gives:

$$T = T_{in} + (T_{out} - T_{in}) \exp\left(-\frac{x}{x_0}\right) \tag{19}$$

with

$$x_0 = \frac{1}{4} \frac{\rho c_p DU}{h_{th}} \tag{20}$$

With U: mean air velocity [ms^{-1}]

Setting $\alpha_{conv} = 0.046$ in the relation of Colburn, the relaxation length becomes:

$$x_0 = 5.43\, Pr^{2/3} Re^{0.2} D \tag{21}$$

By applying equation (18) to a mean diameter of 3 m, which is in the same order of magnitude of that observed in Monlési ice cave, thermal equilibrium is reached after about 700 m. The temperature distribution was modelled with a dataset from Monlési ice cave. Input data was provided by the air temperatures recorded at the base of the entrance shaft during an open period ranging from the 18th to 25th of February 2003. The cave temperatures recorded along the cave roof at 2 m intervals during the same period were inputted for comparison. Due to daily temperature oscillations, heat exchanges are clearly evidenced by a significant smoothing of the temperature variation at increasing distances from the cave entrance (Fig. 5). However, data recorded inside the cave suggests that heat transfer is significantly higher than the modelled values. Therefore, in addition to sensible heat exchanges, major latent heat exchanges must be considered also.

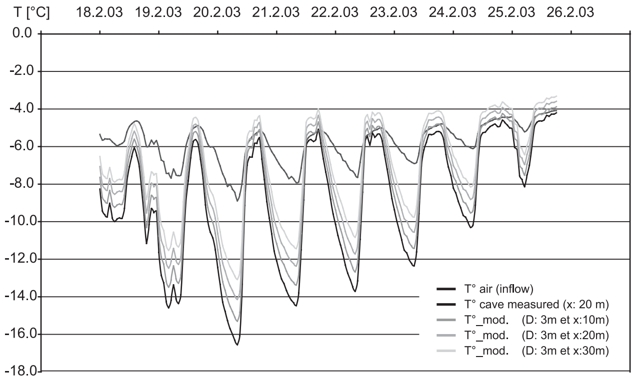

Fig. 5 Modelled air temperatures considering only sensible heat exchanges. Comparison with data recorded inside the cave suggests that latent heat exchanges cannot be neglected.

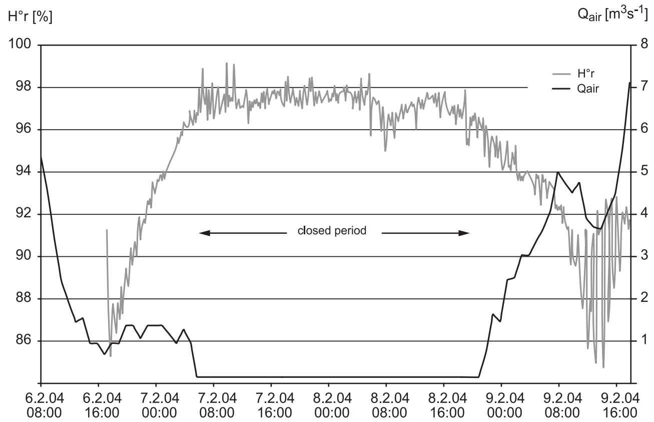

Fig. 6 Humidity of outflowing air. Although data indicate values close to saturation during a closed period, major ventilation events significantly reduce the humidity of cave air.

Model 2

Setting the hypothesis that outflowing cave air is saturated with water, the specific humidity is given by:

$$\omega = m_{dry\_air} \frac{P_v}{P - P_v} \tag{22}$$

$$\omega = \frac{0.62}{(P - Hr \cdot 10^{(5.083-1665.6/(\theta+228))})} \cdot Hr \cdot 10^{(5.083-1665.6/(\theta+228))}$$

Where ω: specific humidity [kg/kg$_{as}$]; m_{dry_air}: mass of dry air [kg]; P_v: partial pressure of vapour [10^5 Pa]; P: air pressure [10^5 Pa]; Hr: relative humidity (1 for saturated air); θ: air temperature [°C].

As the temperature of the outflowing air is known, its enthalpy is computed by applying equation (22) to relation (10). The energy advected out of the sub-system by air circulation reached 22 GJ during the 2002-2003 cycle. Since the energy of inflowing air was assessed at -60 GJ for the same annual cycle (see 3.4, p.4) the total heat exchanged by forced convection can be estimated at 82 GJ. This value can be compared to the sensible heat exchanged if phase changes are disregarded (i.e. 43 GJ). Hence the thermal effect of evaporation/sublimation can be assessed at 39 GJ, which is of the same order as the energy stored by pure sensible heat transfers.

However, Figure 6 shows the humidity of outflowing air measured during a three-day survey in winter 2004. Although the data underlines values near to saturation during a closed period (i.e. Q_{air}: ~0 m^3s^{-1}), it is also shown that ventilation significantly reduced the specific humidity of the cave air. Since the efficiency of sublimation is not maximal, the computed effect of evaporation on the energy balance is considered as an upper value. Thus, an uncertainty of 30% is applied to computed values.

3.5 The thermal contribution of infiltration water

Besides heat transfers induced by air circulation, the defined sub-system also exchanges energy with percolating water. Hence, the thermal contribution of water infiltrations was determined in order to quantify their relative importance in the energy balance.

3.5.1 Actual infiltration

Figure 7 illustrates the water discharge recorded between November 2002 and October 2003 at the main inlet of Monlési ice cave. Maximal values measured during this time-period reached nearly 15 lmin^{-1}. The analysis of 34 flood events concluded that the discharge occurs within a time delay of 1 to 8 hours after the first significant precipitation, while the recession ends within two days after the maximal discharge peak (cf. appendix). Field observations make it possible to set the mean recession time at about 2300 minutes (~39 h). This data suggests a highly transmissive system with a strong control from exterior precipitation distribution.

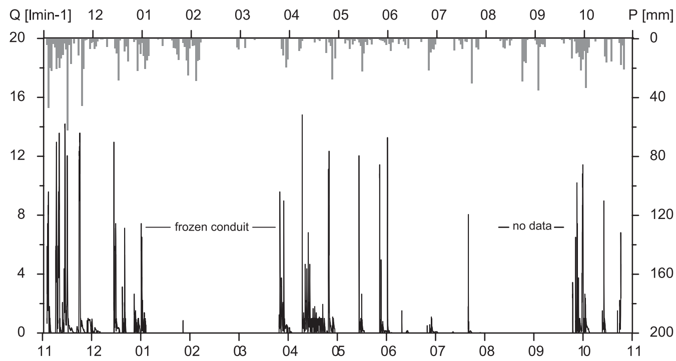

Fig. 7 Pluviometry and water discharge recorded at the main inlet of Monlési ice cave.

The hydrological balance in a given time period is provided by:

$$I=(P-ETR)*CA-\Delta V_{stock} \tag{23}$$

Where I: actual infiltration [l]; P: pluviometry [mm]; ETR: effective evapo-transpiration [mm]; CA: catchment area [m^2]; ΔV_{stock}: water storage in the soil/epikarst aquifer [l].

Evapo-transpiration is minimal when biological activity and air temperatures are low. The soil/epikarst aquifer overlying the cave is assumed to be almost saturated during the autumn as a result of frequent precipitation events at this time of year. According to data from Tripet (1972, p. 72-80), the mean specific infiltration could be assessed at more than 95% of precipitation from November to March. Hence, data from five distinct flood events in November 2002 were used to compute the catchment area of the main water inlet of Monlési ice cave. The comparison of recorded precipitation values with measured discharges suggests a mean catchment area of about $240 \pm 25\,m^2$ (Tab. 1). However, data from 2003 emphasizes the non-stationary relationship between precipitation and measured discharges (Tab. 2). This fluctuation is attributed mostly to the storage capacity of the hydrological system composed of the overlying soil and the epikarst aquifer. As suggested by Jeannin & Grasso (1995), the recharge of this reservoir is more dependent on the duration of precipitation events than on their intensity.
As shown by Table 2, the dense vegetation observed in the catchment area leads to a mean specific infiltration of only ~20% between May and October 2003.

During major flood events this value reaches an average 34% of the precipitation volume, consistent with regional observations (Tripet, 1972). But owing to the runoff observed during short and intensive events like thunderstorms, fast drainage towards karst conduits can represent almost 100% of the volume of precipitations measured within the catchment area.

3.5.2 Reconstructing missing water discharges
In order to compute the heat transferred by water to the air system, missing data were reconstructed for the annual cycle 2002-2003. Although the determination of a reliable transfer function is not possible in a simple approach, a rough approximation of flood events is proposed for the reconstruction of missing data. Based on three distinct short-term events of July 2003, the recession hydrograph of the main water inlet of Monlési ice cave was approximated by:

$$Q_t=(Q_0^{-1})e^{-0.01t} + 1e^{-0.001t_1} \tag{24}$$

Where Q_t: discharge at time t [$lmin^{-1}$]; e: base of the Napierian logarithms (2,718...); Q_0: maximal discharge [$lmin^{-1}$]; t: time elapsed since discharge peak [min].

Empirical observations determined Q_0 as a function of actual infiltration:

$$Q_0=0.0065*P*CA*C - 4.2695 \tag{25}$$

With P: precipitation [mm]; CA: catchment area [m^2]; C: correction factor related to the actual infiltration.

Event	Date		Precipitations [lm^{-2}]	Estimated evapo-transpiration [lm^{-2}]	Measured water volume [l]	Catchment area [m^2]
#1	05.11.2002	15:30	44	3	9328	226
#2	10.11.2002	04:00	38.6	2	9187	253
#3	11.11.2002	21:00	42.6	3	11140	278
#5	17.11.2002	05:00	85.4	5	17837	222
#6	25.11.2002	01:00	64.4	4	13270	219
						240 ± 25

Tab. 1 Computed catchment area of the main water inlet in Monlési ice cave during November 2002. Frequent precipitation events during autumn lead to a water saturated soil/epikarst aquifer and to an insignificant evapotranspiration: actual infiltration was estimated to be of nearly 100%.

Event	Date	Volume [l]	Pluvio [lm^{-2}]	Catchment area $I_{eff.}$ 100% [m^2]	Actual infiltration	CA vs actual infiltration
#1	01.05.2003	5151	36.85	140	58%	233
#2	03.05.2003	310	14.025	22	9%	221
#4	21.05.2003	1970	49.06	40	17%	201
#6	02.06.2003	2797	10.065	278	116%	278
#7	06.06.2003	1004	15.07	67	28%	222
#8	15.06.2003	45	12	4	2%	250
#11	04.07.2003	104	6.8	15	6%	218
#14	27.07.2003	432	32	13	6%	225
#20	09.10.2003	15646	147.8	106	44%	212
#22	03.11.2003	5379	53.2	101	42%	225
#23	13.11.2003	60	10.9	6	2%	222
#24	17.11.2003	1650	18.8	88	37%	219
#25	30.11.2003	117	45	3	1%	259
	Total	**34662**	**452**	**77**	**34%**	**230**

Tab. 2 Computed catchment area of the main water inlet in Monlési ice cave in 2003. Data emphasizes a non-stationary relation between precipitation and measured discharges.

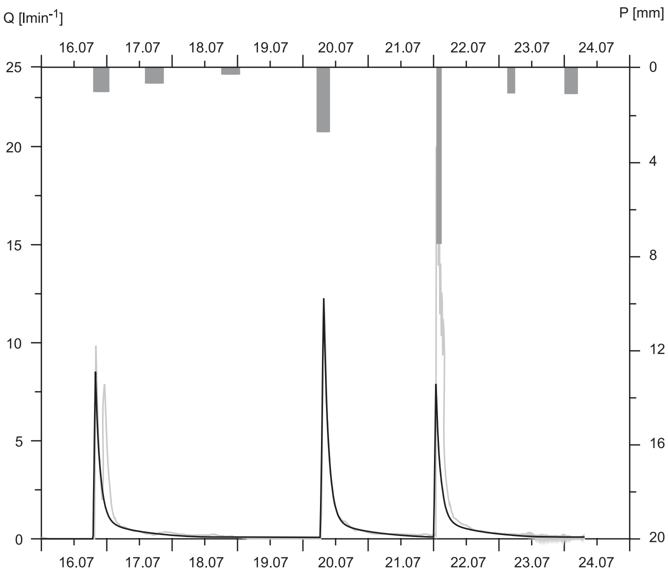

Fig. 8 Reconstructed water discharge at the main inlet in Monlési ice cave. Major uncertainties remain on the reconstructed precipitation amounts, which tend to be systematically underestimated during thunderstorms.

Figure 8 illustrates reconstructed discharges from three distinct events. Accuracy of computed water volumes is assessed at ± 30%. Two of the modelled recession curves fit well with measured values but major uncertainties remain on the reconstructed precipitation amounts which tend to be systematically underestimated during thunderstorms (for instance, peak #3, Fig. 9). Nevertheless, this approach was assumed to be valuable enough for the reconstruction of water discharges at the main inlet of Monlési ice cave.

3.5.3 Discharge versus water temperature

According to the recorded data, the main water inlet remains frozen about 40% of the year (3411 hours from 11.2002 to 11.2003). Measured water temperatures were analysed from June to October 2003, and Figure 9 outlines the strong correlation observed between temperature and the measured discharges. Maximal temperature values are reached with high discharges, when heat exchange with the surrounding rock is minimal. Those correspond to the temperature at the system's boundary conditions within the limestone (i.e. exterior MAAT: 4.5 °C). Although several datasets are missing for a physically-based modelling, the relation between the temperature and the discharge could be determined empirically as:

$$T_{water} = 0.17 \cdot Q - e^{(-2.5Q+0.6)} + 1.6 \quad (26)$$

Where T_{water}: water temperature [°C]; Q: discharge [$lmin^{-1}$]; e: base of the Napierian logarithms (2,718...).

Since modelled temperature curves fit relatively well with measured values (r^2: 0.9), the accuracy of this reconstruction can be considered as good. However, values are reliable only for flood events presenting a similar signature (i.e. short intensive events), which was actually the case during the six weeks of missing data.

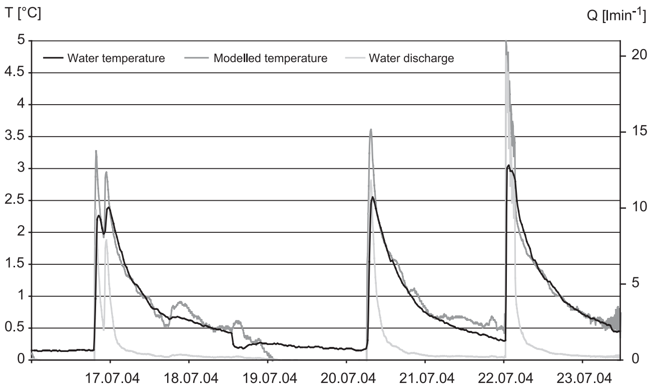

Fig. 9 Reconstructed water temperatures during flow events in Monlési ice cave. Maximal values are reached with high discharges when heat exchange with the surrounding rock is minimal.

3.5.4 Energy supplied by water infiltrations

Knowing the temperature and the discharge at any time at the main water inlet, the energy supplied to the system is given by:

$$P = \frac{c_w \int_{t1}^{t2} Q_m \theta dt}{t_2 - t_1} \quad (27)$$

Where P: mean power of inflowing water [W]; c_w: heat capacity of water; Q_m: water flow mass rate [kgs^{-1}]; θ: temperature of water [°C]; t_1, t_2: time at instant 1 and 2 respectively.

Setting the hypothesis that the heat exchange with the cave air is maximal, it was assumed that the outflowing water temperature is at 0 °C. Hence, the thermal contribution of the main water inlet during the 2002-2003 annual cycle corresponds to 2.8 GJ. Since it is known that this inlet drains about half of the water circulating through the cave, the total heat transferred to the cave by water infiltrations can be assessed at about 5.6 GJ during the observation period. Accuracy of this value is estimated at ± 20%.

3.3.5 Discussion

By monitoring the discharge and the temperature at the main water inlet, it was possible to assess the contribution of water infiltrations to the energy balance of the studied system. It was shown that the analysis of single infiltration events enabled the reconstruction of flood peaks under similar hydrological conditions (for instance thunderstorms). But, by demonstrating that the relation between precipitation and discharge is non-stationary, the modelling of actual infiltration during a complete annual cycle is almost impossible with a simple approach. A consistent relationship between measured precipitations and actual infiltration is achieved only by providing an extensive model of evapo-transpiration and water storage within the soil-epikarst aquifer. This approach allows the energy balance of the ice cave (i.e. the mass balance) to be linked to precipitation measurements issued from nearby MeteoSwiss stations.
Furthermore, assuming that the heat exchange is maximal (i.e. outflowing water temperature at 0 °C) has to be considered as a little oversimplification of our model. In fact, the boundaries of the global system were defined by a constant rock temperature of 4.5 °C suggesting new heat exchanges with the surrounding limestone. Therefore, it is very likely that the thermal contribution of infiltration water is slightly overestimated. But since the computed heat supply demonstrated that the contribution of water infiltrations to the energy balance of the cave is small, a more accurate infiltration model appears to be superfluous for our applications.

3.6 Energy balance of the sub-system air

As defined in Chapter 3.2, the energy balance of the sub-system air is given by:

$$\Delta H = (E_{air} - E'_{air}) + (LE_{cond.} - LE_{sub.}) + (A_{water} - A'_{water}) \quad (28)$$

Where ΔH: energy lost by the sub-system; E_{air}: sensible heat advected by air into the system; E'_{air}: sensible heat advected by air out of the system; $LE_{cond.}$: latent heat related to condensation; LE_{sub}: latent heat related to sublimation; A_{water}: heat advected by water into the system; A'_{water}: heat advected by water out of the system.

Measurements performed during the annual cycle 2002-2003 provided the following orders of magnitude for each of these terms:

ΔH= (-142.8 GJ − (-99.7 GJ)) + (7 GJ -38.7 GJ) + (5.6 GJ - 0 GJ)
ΔH= **-69.2 GJ**

Hence, the energy lost by the sub-system air in favour of the rock or the ice (ΔH) can be estimated at **-69.2 GJ** during the annual cycle 2002-2003, with an accuracy of ±30%. This energy might be stored in the form of sensible heat in the surrounding rock or ice, or used for the freezing of infiltration water.

3.7 Discussion

By quantifying the energy balance of the sub-system air, one can conclude that heat transfers related to evapo-condensation effects cannot be neglected. Although they were assumed to be almost as important as heat transfers induced by sensible heat exchanges, their quantification relied on the assumption that cave air remains close to saturation ($H°_r$=100%). However, sporadic measurements demonstrated that this was not the case during major ventilation events.
It was shown that air circulation is controlled by the temperature difference between the cave air and the exterior atmosphere. This difference is maximal on cold winter days, mainly in continental meteorological contexts. Since these periods are also associated with dry air masses, evaporation is assumed to be significant within the cave. But owing to the short distance within the conduit, this process is not sufficient to expect an outflowing air saturated in water. The estimated value of 38.7 GJ, related to evaporation rates during the annual cycle 2002-2003, therefore has to be considered as an upper approximation.
Although major ventilation events are at the origin of an unbalanced energy budget, it was shown that the efficiency of heat transfers is improved with slow air velocities. Hence, for the same air volume transiting through the cave, the observed heat exchange (i.e. the cooling of the system) is increased with the duration of the open system.

The energy supply of water infiltrations appears to be an order of magnitude lower than heat transfers related to air circulation. Thus, it can be assumed that modifications in the annual discharge rate play only a minor role on the final energy balance.

4 | The sub-system ice

4.1 Introduction

Borehole thermometry is a tool for predicting the temperature distribution within an ice filling. This latter deserves attention both for the role played in the global energy balance of an ice cave and its relation with potential paleoclimatic records.
Measured temperature profiles were used to validate empirically the heat diffusion model within the ice filling of Monlési ice cave. Applying an iterative process, 1D and 2D modelling aimed at reconstruction of the temperature distribution in order to assess the sensible heat stored within the ice.

4.2 Definition of the sub-system

The investigated sub-system is composed of a well defined ice (s.l.) body (Fig. 10). The system is assumed to have a constant volume and heat transfer occurs only by diffusion. Boundary conditions are given by daily and seasonal temperature fluctuation at the upper and lateral limits, while a constant temperature of 0 °C is assumed at the lower limit.

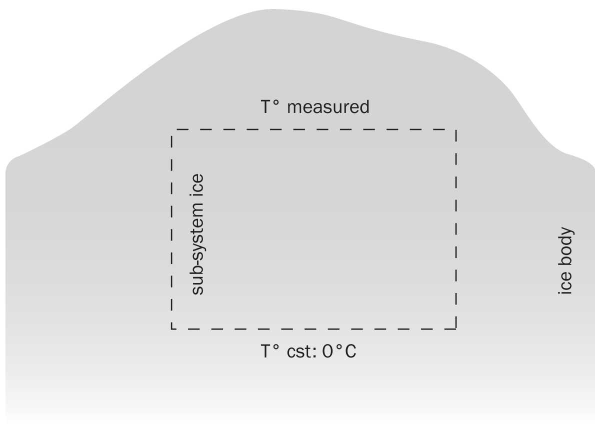

Fig. 10 Delineation of the sub-system ice. Boundary conditions are given by daily and seasonal variations at the upper and lateral limits, while a constant temperature of 0 °C is assumed at the lower limit.

4.3 The thermal properties of the ice

Ice accumulation in Monlési ice cave results chiefly from refreezing of infiltration water, but firn accumulations can be observed at the base of the entrance shafts. Density measurements carried out on five ice samples show a mean value of about 970 kgm^{-3} (Tab. 3). This data suggests that the filling contains mostly congelation ice and that thermal properties of pure ice might be considered.

Paterson (1994) provided a temperature-dependent relation for the determination of the specific heat capacity of ice:

$$c_i = 152.5+7.1222*T \tag{29}$$

Where c_i: specific heat capacity of ice [$Jkg^{-1}K^{-1}$]; T: temperature [K].

At temperatures close to the melting point, the computed heat capacity is about 2090 $Jkg^{-1}K^{-1}$. The thermal conductivity (λ) of pure ice can be determined by the empirical formula provided by Paterson (1994):

$$\lambda=9.828*e^{(-5.7*10^{-3}*T)} \tag{30}$$

Therefore, the heat conductivity of pure ice can be assessed at 2.1 $Wm^{-1}K^{-1}$ at temperatures close to the melting point. The thermal conductivity is related to the heat diffusivity by the relation:

$$\lambda = a*\rho*c \tag{31}$$

Where λ: heat conductivity [$Wm^{-1}K^{-1}$]; a: heat diffusivity [m^2s^{-1}]; ρ: density [kgm^{-3}]; c: specific heat capacity [$Jkg^{-1}K^{-1}$].

Validation of these theoretical values is provided by temperature measurements carried out in a borehole.

Volume [ml]	**Mass** [g]	**Density** [m^2]
210	192	0.91
105	104	0.99
235	220	0.94
250	246	0.98
250	244	0.98
120	120	1.00
		0.97 ± 0.05

Tab. 3 Density of cave ice measured in Monlési cave. Despite the low accuracy of measurements, data suggests the presence of pure congelation ice.

4.4 Data acquisition

Steam drilling was performed in Monlési ice cave in order to estimate the temperature distribution inside the ice filling. A light-weight system (Heucke, 1999), typically used for firn drilling projects, was requested. Easily transportable in the field, this system has a water tank of about four litres and a power alimentation provided by two small butane-propane gas cylinders (approx. 320 gh^{-1}). The steam drilling was performed with a 34 mm jet, and an operating pressure varying between 0.8 and 1.6 bar providing borehole diameters of approximately 40 mm. Due to a pressure drop induced by the meltwater column, the expected investigation depth is about 20 m in massive congelation ice.

Morphological evidence in the cave suggested an ice thickness ranging between 10-15 m (Luetscher & Wenger, 2002). Most of the 22 drillings attempted during the February 2002 campaign could not drill deeper than about 1 m due to the presence of numerous small cryoclasts and ice impurities (wooden particles, stones, etc.). Therefore, an accurate estimate of the ice thickness is lacking, but fortunately, one of the drills provided an 8.5 m deep borehole before it was stopped by some wood inclusion. The drainage of borehole water at a 5 m depth also suggested the presence of a major discontinuity interpreted as a fracture.

This «deep» borehole was equipped with six Pt100 thermistors, three of which were previously calibrated in an ice mixture bath. Final accuracy is assessed at 0.1 °C, but resolution is given at 0.001 °C. Thermistors were positioned between 0.1 and 8 m depth on February 28th 2002. Due to in-borehole meltwater, the drill refroze the following day.

4.5 Results

Ice temperatures were recorded during the 2002-2003 annual cycle in 30 minute time steps, but technical problems with the logger caused a major gap in the time series from 9.08.03 - 25.09.03. Furthermore, due to the melting of the ice surface, the thermistor placed at 0.1 m depth was buried out of the ice body in October 2002 and broke.

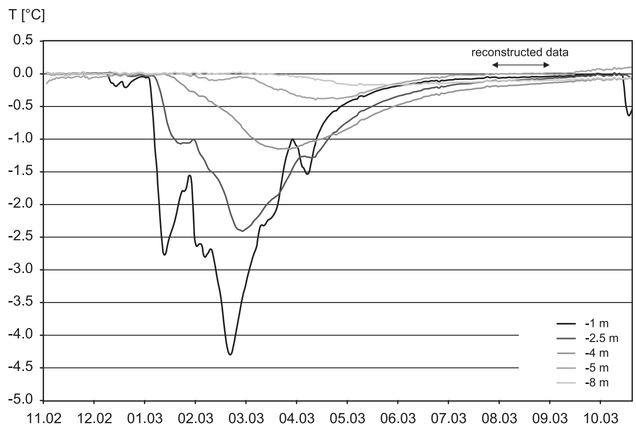

Fig. 11 Daily mean temperature recorded at different depths in the ice filling of Monlési cave. Seasonal fluctuations are observed over the entire filling. Data suggests a temperate ice body.

Figure 11 illustrates calibrated ice temperatures recorded from November 2002 to October 2003 at -1, -2.5, -4, -5 and -8 m depth. Seasonal oscillations are observed over the entire length of the borehole but temperatures recorded at -8 m outline a mean annual value close to 0 °C suggesting a temperate ice filling (Paterson, 1994). Thus, one can conclude that the temperature at rock-ice interface is 0 °C suggesting that melting processes affect the ice filling from all sides. This assumption is validated by field observations of accessible parts along the ice filling.

4.5.1 Thermal diffusivity of the ice

The hypothesis proposes a homogeneous ice body (i.e. heat diffusivity=constant) with no internal heat generation. Considering a one-dimensional heat conduction, the temperature at a depth z is given by the heat equation:

$$\frac{\partial T}{\partial t} = a\frac{\partial^2 T}{\partial z^2} \tag{32}$$

Where T: Temperature; t: time; a: thermal diffusivity; z: depth.

Equation (32) is solved numerically using a finite difference scheme (e.g. Mitchell & Griffiths, 1987), where derivatives are replaced by differences. A point i at the depth z=Δz(i+1) is surrounded by two points i-1 and i+1. T is the temperature at a given instant t and T' corresponds to the temperature at t+Δt.

$$\frac{\partial T}{\partial t} \Rightarrow \frac{{}^2T}{{}^2t} = \frac{T'_i - T_i}{2^2 t} \tag{33}$$

$$\frac{\partial^2 T}{\partial z^2} \Rightarrow \frac{\partial}{\partial z}\left(\frac{{}^2T}{{}^2z}\right) \approx \frac{(T_{i+1}-T_i)-(T_i-T_{i-1})}{{}^2z^{\ 2}} = \frac{T_{i+1}-2T_i+T_{i-1}}{{}^2z^{\ 2}} \tag{34}$$

and where equation (33) becomes:

$$\frac{T'_i - T_i}{2^2 t} = a\frac{T_{i+1} - 2T_i + T_{i-1}}{{}^2z^{\ 2}} \tag{35}$$

This equation enables determination of the temperature of a point i at an instant t+dt if the temperature is known at the points i-1, i and i+1 at the instant t:

$$T'_i = T_i + a\,\frac{2^2 t}{{}^2z^{\ 2}}\,(T_{i+1} - 2T_i + T_{i-1}) \tag{36}$$

For a simplification of the computing, a mesh increment in accordance with relation (37) is recommended (Schmidt Method):

$$a\frac{2^2 t}{{}^2z^{\ 2}} = 0.5 \tag{37}$$

where equation (36) gives:

$$T'_i = \frac{1}{2}\left(T_{i+1} + T_{i-1}\right) \tag{38}$$

This approach was applied to the temperature time series shown in Figure 11. The thermal diffusivity of the ice was set at $1*10^{-6}\ m^2s^{-1}$, in agreement with values found in the handbooks. In order to avoid effects of convective heat transfers, the input was taken as the temperature measured within the ice at -1 m. As shown on Figure 12, modelled values fit perfectly with data measured at -2.5 m (r^2: 0.99). But deeper, an offset is observed between the measured values and the modelled curves. This offset is attributed to the simplification introduced by the one-dimensional model. Since the borehole being drilled is at the top of a concave ice volume, this assumption is valuable for a reduced ice thickness only and lateral effects have to be considered with increasing depth. Two-dimensional modelling was attempted in order to improve the model of heat diffusion within the ice mass. The 2D heat equation is given by:

$$\frac{\partial T}{\partial t} = a\left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2}\right) \tag{39}$$

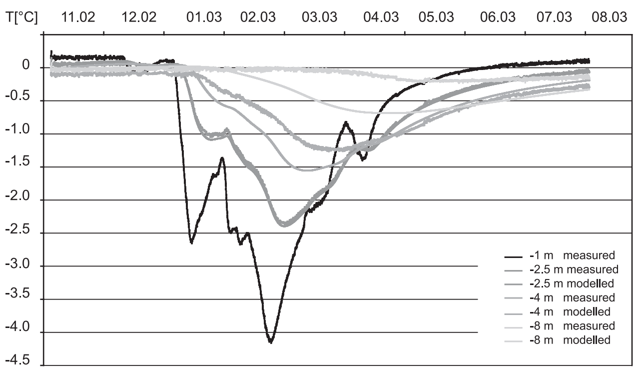

Fig. 12 Comparison between the recorded ice temperatures and the modelled 1D heat diffusion. Data suggests a 1D thermal diffusion within the first metre but lateral effects have to be considered deeper.

Therefore, an approached geometry of the filling was proposed (Fig. 13) and the heat diffusion equation (39) was solved again with a finite difference scheme, where:

$$T'i, j = \frac{T_{i+1,j} + T_{i-1,j} + T_{i,j+1} + T_{i,j-1}}{4} \tag{40}$$

Figure 14 shows that modelled values fit better with recorded time series. Thus, the theoretical values of thermal diffusivity can be considered as consistent with field observations. Hence, the assumption can be set that the ice filling is a homogeneous body composed of pure ice. However, a singular discrepancy between modelled values and measured data could be observed for the thermistor set at -5 m. This difference is attributed to a local discontinuity of the ice body induced by a fracture. Its presence was confirmed during the drilling by the drainage of the borehole-water.

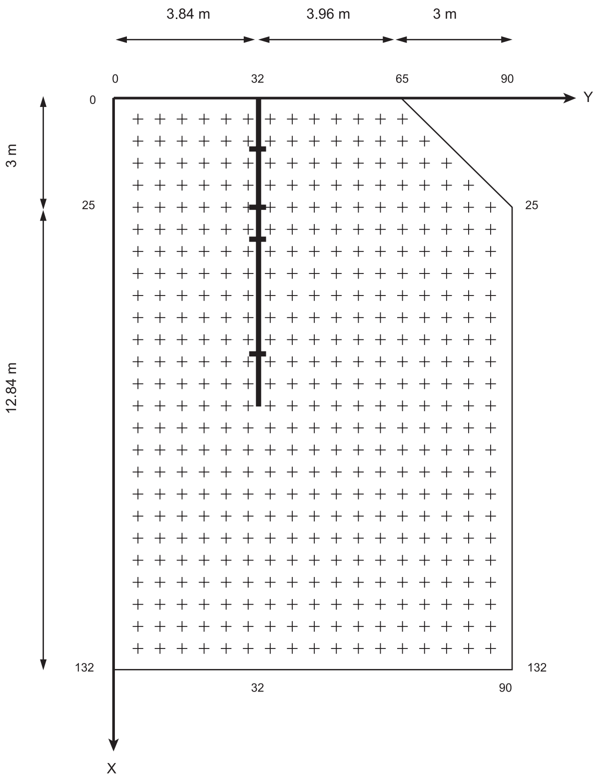

Fig. 13 Approached geometry of the ice filling used for 2D-modelling of heat diffusion. According to the Schmidt method, the mesh increment corresponds to 12 cm (time-step of 3600 s.). Boundary conditions at the top and right-hand side of the model are given by measured temperature oscillations. On the bottom and the left-hand side heat flux is considered as insignificant and is thus taken as zero.

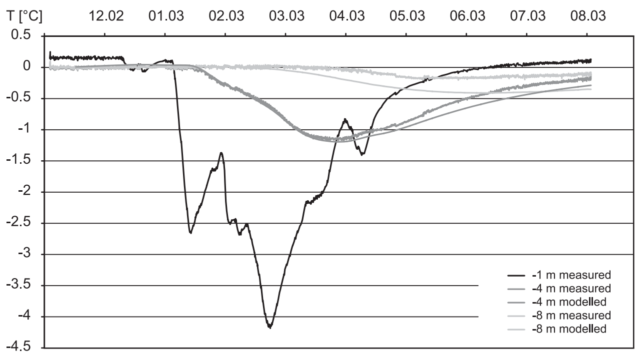

Fig. 14 Comparison between the recorded ice temperature and the modelled 2D heat diffusion. The 2D-modelling significantly improves the reconstructed temperature record.

4.5.2 The sensible heat stored within the ice

The energy employed for raising the temperature of a body is called «sensible heat». According to the heat capacity of the considered material, this energy might be stored in more or less large amounts in a given volume. From morphological evidence, the ice volume was estimated at 6000 m^3 (Luetscher & Wenger, 2002) while the heat capacity of the ice was set at 2040 $Jkg^{-1}K^{-1}$. Since the geometry of the ice filling cannot be approximated with sufficient accuracy yet, the recorded temperatures were considered as representative of the entire ice filling. Computing the sensible heat storage was achieved by discretizing the body into smaller elements, so that the measured temperatures could be attributed to each of them. The sensible heat stored at a given instant was assessed by summing these elements over the entire ice volume.

Hence, the maximal sensible heat stored within the ice during the 2002-2003 cycle could be assessed at **-11.6 GJ** with an uncertainty estimated at ± 30%.

4.5.3 Thermal contribution of intrusive snow accumulations

According to daily measurements performed in La Chaux-de-Fonds (pers. comm. Travaux Publics), cumulated snow precipitation during the winter 2002-2003 reached nearly 2 m thickness. Setting the hypothesis that this value is representative of the Monlési site and assuming a mean snow density of about 300 kgm^{-3}, the snow mass accumulated at the base of the entrance shafts of Monlési ice cave (i.e. 140 m^2) could be assessed at 84 tons. Thus, the latent energy supplied by intrusive snow accumulations could be assessed at **-28 GJ** during the 2002-2003 observation periods. Since a major uncertainty remains on the snow density, final accuracy is estimated at ± 30%.

4.6 Discussion

Computed data validates the heat diffusion model applied to the ice filling of Monlési ice cave. Hence, the ice filling can be considered as almost homogeneous on a large scale. Thermal diffusion within the ice is controlled by 1D conduction within the first metres. However, at greater depth (i.e. below ~2.5 m) 2D heat diffusion must be considered. These models confirmed that a thermal diffusivity of $a_{ice}=1*10^{-06}$ m^2s^{-1} seems to be adapted to our purpose. The heat conductivity was assessed at $\lambda=2.02$ $Wm^{-1}K^{-1}$.

The 2D approach significantly improved the reconstructed temperature distribution within the ice filling. This model could still be improved by a better approximation of the ice geometry and by a 3D finite-element model. But at present, uncertainties on the extension of the ice filling are too important to justify this further step.

Since it was shown that the ice body already reaches a thermal equilibrium in mid August, it can be assumed that sensible heat storage is not relevant for the annual energy balance of the ice cave. However, the sensible heat stored by the ice body tends to reduce the duration of the melt period.

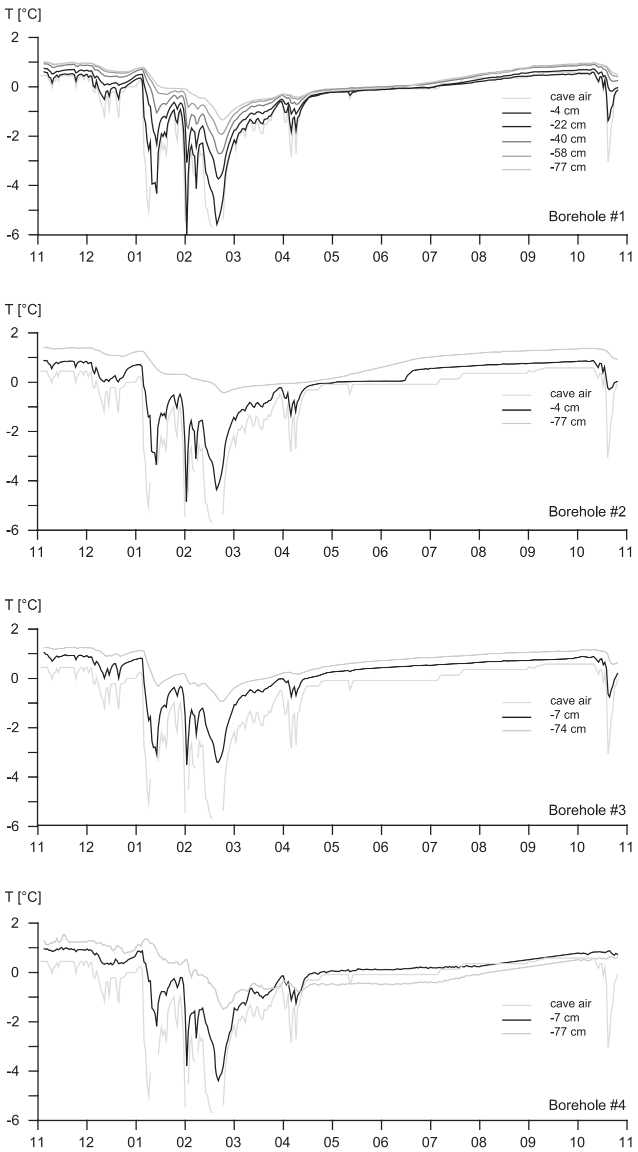

Fig. 15 Rock temperatures measured in four different boreholes of Monlési ice cave between 11.2002 and 10.2003.

5 | The sub-system rock

5.1 Introduction

The temperature distribution in karst systems has been shown to be controlled by air and water fluxes (Luetscher & Jeannin, 2004; Part II, p. 52). Subsequently, the temperature at a given point of the karst massif is assumed to correspond to the exterior Mean Annual Air Temperature (MAAT) at the same altitude. The thermal anomaly induced by an ice cave leads to significant temperature gradients in the surrounding rock. Thus, major heat exchanges can be expected between the limestone and the cave. In order to quantify this ground heat flux, borehole temperatures were recorded during an annual cycle in rock walls of Monlési ice cave. Results presented below aim at determining the distance of the boundary condition, the mean ground heat flux, and the sensible heat stored within the limestone after winter cooling of the cave.

5.2 System description

The investigated sub-system is composed of massive limestone with a constant volume. Its boundary condition is geometrically well defined on one side by the cave. On the side of the massif the sub-system is delimited by a constant temperature over time. This temperature is taken as the exterior Mean Annual Air Temperature (MAAT), which is about 4.5 °C at the Monlési site. Thermal characteristics of limestone considered here are issued from handbooks. The thermal diffusivity of the limestone is taken at $1.13*10^{-6} m^2s^{-1}$.

5.3 Data acquisition

Four distinct boreholes were drilled into the cave walls up to a depth of 80 cm. Pt-100 thermistors were set at different depths, isolated between each other by polyurethane packers. In order to improve the heat conduction with the surrounding limestone, each probe was set previously into a copper ring of a diameter close in size to the borehole (i.e. 40 mm). Rock temperatures were recorded from 11.2002 to 10.2003 in 30-minute time intervals using a DT500 logging unit powered by a 6V battery.

5.4 Results

An almost constant temperature gradient was observed at the end of the summer season due to a cave temperature controlled by the melting ice (i.e. 0 °C). Assuming a heat conductivity of the limestone of 2.2 $Wm^{-1}K^{-1}$, the «normal» ground heat flux could be determined by:

$$\varphi = \lambda \frac{\Delta T}{\Delta x} \qquad (41)$$

Where φ: density of the ground heat flux [Wm^{-2}]; λ: heat conductivity [$Wm^{-1}K^{-1}$]; ΔT: temperature difference [K]; Δx: distance between the two measurement points [m].

Borehole	Observation period	ΔX [m]	ΔT [°C]	Gradient [°Cm^{-1}]	Heat flux [Wm^{2}]
#1	25.09.03–05.10.03	0.73	0.38	0.52	1.1
#2	25.09.03–05.10.03	0.73	0.51	0.70	1.5
#2	01.08.04–08.08.04	0.73	0.50	0.68	1.5
#3	25.09.03–05.10.03	0.67	0.31	0.46	1.0
#3	01.08.04–08.08.04	0.67	0.32	0.48	1.1
#4	04.11.02–15.11.02	0.7	0.31	0.44	1.0

Tab. 4 Heat fluxes measured in four boreholes drilled in rock walls of Monlési cave.

During autumn, observed temperature gradients varied between 0.5 °C/m in borehole #3 and 0.7 °C/m in borehole #2, suggesting a ground heat flux ranging between 1 and 1.5 Wm^{-2} under steady state conditions (Tab. 4). This estimation is consistent with interpretations from a multi-annual topometric survey which concluded that there is a basal melting of nearly 10 cmyear^{-1} (Luetscher, 2004). Therefore, setting the hypothesis of a homogeneous massive limestone (λ: 2.2 Wm^{-1}K^{-1}), equation (39) allows for the calculation that boundary conditions (i.e. constant temperature of 4.5 °C) can be found at a distance ranging between 6 to 9 m from the cave walls.
Measured rock temperatures presented on Figure 15 show a good correlation with cave air temperature fluctuations, suggesting significant heat exchanges between both sub-systems. The progressive damping of the signal with increasing depth was interpreted to be related to the heat diffusion. In order to validate the diffusion model, the equation of the assumed 1D heat conduction was solved in a finite-difference scheme (e.g. 4.5.1). However, reconstructed data concluded that the apparent heat diffusion within the limestone is significantly lower than expected and ranges between 0.3 and 0.5*10^{-6} m^2/s (instead of $a_{limestone}$: 1.13*10^{-6} m^2/s).
This low value was attributed to latent heat effects in the first stage. Phase changes of pore water are well illustrated by the abrupt temperature change observed after a long stable period corresponding to freezing/thawing events (i.e. 0 °C; Fig. 16). But, the irregular temperature gradients observed between five thermistors set in borehole #1 also suggest that advective heat transfers cannot be neglected. Actually, temperature data from this borehole suggested the presence of a cold air/water flow at about 50 cm from the cave walls.
Since the limestone was assumed to be too heterogeneous to be the object of an accurate modelling of heat transfers, the sensible heat storage was assessed empirically by discretizing the rock wall into elements associated with the recorded data. Heat capacity of limestone was taken at 810 Jkg^{-1}K^{-1} and the rock density was measured as 2.6. The sensible heat storage was determined for a given rock volume extending from the air-rock interface to the boundary condition of constant temperature.
Setting the hypothesis that borehole #1 is representative of the rock temperature distribution within the cave and assuming a rock-air exchange surface of about 1100 m^2, the maximal sensible heat stored during the 2002-2003 was assessed at -9.2 GJ with an uncertainty estimated at ± 30%.

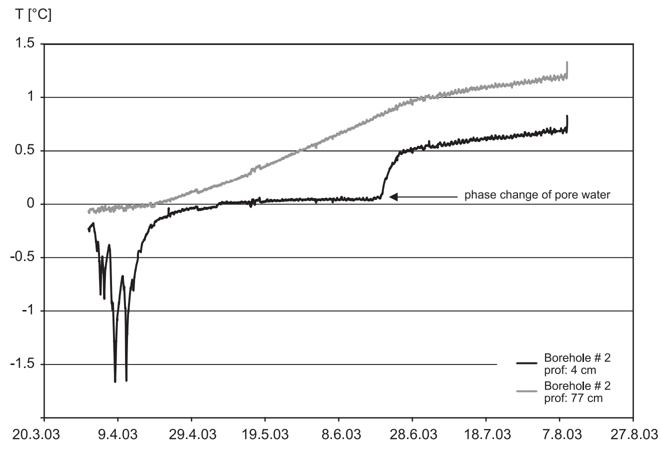

Fig. 16 Rock temperature recorded from April to July 2003 in borehole #2. The thawing of the rock is evidenced by the phase change of pore water.

Setting the hypothesis that a mean constant ground heat flux of 1 Wm^{-2} is representative of the defined sub-system, the energy supplied by the massif is given by:

$$E_{rock} = \left(\int_{time} \varphi S dt\right) - E_{sens} \qquad (42)$$

Where E_{rock}: energy issued from the ground heat flux [J]; φ: density of heat flux [Wm^{-2}]; S: exchange surface; E_{sens}: Sensible heat stored within the rock [J].

From (41), and considering an exchange surface of 4000 m^2 corresponding to the rock-air and rock-ice interfaces, the energy issued from the sub-system rock (E_{rock}) was assessed at **117 GJ** for the 2002-2003 observation period.

5.5 Discussion

Since the analysis of borehole temperature gradients confirmed the presence of air and/or water circulations in Monlési rock walls, the limestone surrounding Monlési ice cave cannot be considered as homogeneous. These advected fluxes are assumed to increase significantly the observed heat exchanges with the cave. In fact, under steady-state conditions temperature gradients up to 0.33 °C/36 cm (i.e. φ : ~2 Wm^{-2}) could be observed locally (e.g. Borehole #1). Evidence of phase changes induced by freezing temperatures was supported by the abrupt temperature change observed after long stable periods. Data from Borehole #4 (Fig. 15) also suggested that gradient inversions could occur locally, counterbalancing the heat supplied by ground heat fluxes. However, data from the other boreholes suggest that this process is limited to the entrance zone of the cave.
By assuming a mean heat flux of 1 Wm^{-2} issued from the cave surroundings, the energy supplied by conductive heat exchanges could be assessed at **117 GJ** during the 2002-2003 annual cycle. However, major doubts remain on the representativeness of the considered heat flux. Furthermore, the estimation of the exchange surface and of the sensible heat stored during winter cooling is subject to significant uncertainties. Although this approximation has its limitations, final accuracy of the estimated energy supply was assessed at ± 50%.

6 | Energy balance of the Monlési ice cave

As defined in Chapter 2, the annual energy balance of Monlési ice cave has been defined as:

$$\Delta E_{air} + LE_{snow} + LE_{ice} = S + \Delta E_{water} + R \tag{43}$$

Where ΔE_{air}: sensible and latent heat related to the advected air flow; LE_{snow}: latent heat of intrusive snow; LE_{ice}: latent heat of ice; S: ground heat flux; ΔE_{water}: heat advected by water circulation; R= solar radiation.

Owing to the cave morphology and the surrounding topography, direct solar radiation is almost never observed in the entrance pits. The hypothesis was set that the diffuse radiative heat flux could be neglected due to the presence of dense vegetation at the shaft entrances. This simplification is supported further by the high albedo of the snow accumulated at the base of the entrance shafts. Hence, heat transfers observed during the annual cycle 2002-2003 enable quantification of the energy balance of Monlési ice cave as following:

$$75\ \text{GJ} + 28\ \text{GJ} + LE_{ice} = 117\ \text{GJ} + 6\ \text{GJ} + 0\ \text{GJ}$$

Setting the hypothesis that the disequilibrium of the energy balance is completely compensated for by the melting or freezing of cave ice, the adjusting parameter LE_{ice} can be assessed at an energy of 20 GJ during the annual cycle 2002-2003. This energy corresponds to the disappearance of an 8 cm thick ice layer which is close to observations performed on the fluctuation of the extent of cave ice volume (Fig. 17).

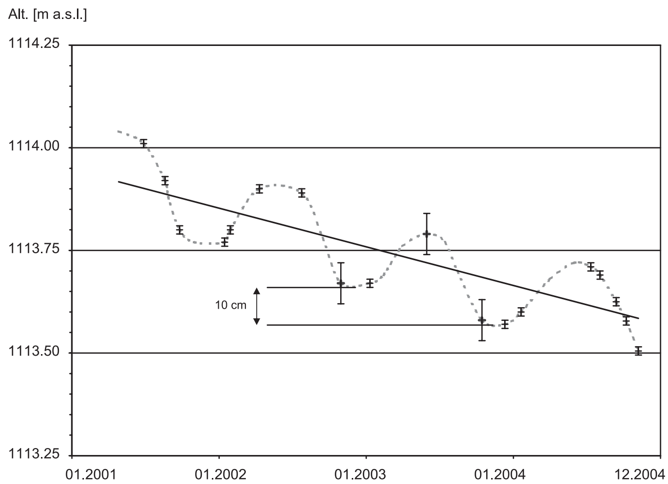

Fig. 17 Ice fluctuations measured in Monlési ice cave between 2001 and 2004. Vertical bars reflect the accuracy of measurements. The thickness of the ice body decreased about 10 cm during the 2002-2003 annual cycle.

7 | Discussion

The analysis of the energy balance of Monlési ice cave clearly shows that fluctuations of the ice volume should reflect the disequilibrium between the energy loss («cooling») and the energy supply («warming»). Although numerous uncertainties remain on each term of the energy balance, accuracy can be considered as sufficient to highlight the relative importance of the different parameters. The quantification of heat transfers observed in the Monlési ice cave system clearly show that the cave ice mass balance is closely related to: 1) the winter cooling induced by air circulation (forced convection) which is responsible for about 70% of the heat loss; and 2) the latent heat supplied by intrusive snow accumulations, which is responsible for about 20% of the heat loss.

Field measurements demonstrate that the energy related to phase changes induced by air circulation is in the same order of magnitude as pure sensible heat transfers.

Sublimation of cave ice is favoured by the inflow of dry air which is often related to low exterior air temperatures (i.e. high air flows). However, it should be noted that the efficiency of heat transfers decreases with increasing air flow and thus, sensible heat storage is maximal during long periods with small temperature differences between the cave air and the exterior atmosphere rather than during brief cold events.

Ice crystallisation occurs only if sufficient water is available. Thus, the most favourable conditions for ice mass growth are when the temperature outside the cave is above 0 °C (allowing the percolation of snow meltwater or rainwater), while the cave air temperature remains below the freezing point. These conditions are reached preferentially during the early spring when cold nights follow warmer days. But mid spells during the winter seasons also constitute a favourable context for the crystallisation of high amounts of congelation ice.

Conversely to intrusive snow accumulations, the energy supplied by water infiltrations is considered only a minor contribution to the overall energy balance. Thus, it could be demonstrated that the melting of the cave ice is related mostly to heat exchanges with the surrounding limestone. Assuming that boundary conditions within the limestone are defined by a constant temperature, it can be inferred that, due to the high inertia of the rock, the melting rate remains almost constant on a decadal time scale. This interpretation is validated by field observations performed on the cave ice volume, demonstrating the absence of any exceptional ice decrease during the hot summer observed in 2003. However, on a century time scale, temperature fluctuations will modify this boundary condition. It could be shown that in a warming climate context, ground heat fluxes increase according to the long-term trend of MAAT. Depending on the year, more optimal conditions can be reached and the cooling of the ice cave could lead to a cumulated ice formation of nearly 50% more than observed during the 2002-2003 period. If so, an equilibrated cave ice mass balance theoretically could be observed with a mean ground heat flux of about 1.2 Wm^{-2}. This would correspond to a constant boundary condition of nearly 5.4 °C, which is nearly 1 °C higher than in 2002-2003! However, as suggested by photographic comparisons, the cave ice mass balance has been almost negative since 1989 as a result of mild winter temperatures and reduced snow precipitations.

Besides major uncertainties regarding each of the energy balance variables, the primary limitation of this model lies in its confinement to one singular ice cave. Since several morphological parameters must be reconsidered for each cave (e.g. dimensions of the entrance pits and exchange surface with the cave air) this energy balance cannot be easily extrapolated to other ice caves. These concerns notwithstanding, the main conclusions issued from this case-study are validated by the general decrease of cave ice volumes observed during the last fifteen years in the Jura Mountains (e.g. Luetscher et al., 2005). This well documented trend is consistent also with mild winter temperatures and reduced snow accumulations observed since 1989.

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Neuchâtel

Variations spatio-temporelles du volume de glace à la Glacière de Monlési (Boveresse/NE)

*par Marc Luetscher, ISSKA**

**Les mesures effectuées à la glacière de Monlési entre 2001 et 2004 mettent en évidence une forte variabilité saisonnière du volume de glace. Ces observations confirment ainsi que la glace résulte de processus d'accumulation actuels et ne peut être considérée comme une relique de la dernière Glaciation. Néanmoins, un déséquilibre entre le volume de glace formé au cours de l'hiver et le volume fondu annuellement, induit un bilan de masse négatif durant la période couverte par cette étude. Le levé topométrique de quelques stations de références indique un fluage de quelques dizaines de centimètres. Sa composante verticale est attribuée à la fonte provoquée par le flux de chaleur issu du système karstique sous-jacent; sa composante horizontale résulte quant à elle de la fonte au contact des parois.
En vue de mieux établir le lien entre les fluctuations volumétriques de ce remplissage de glace et les changements climatiques extérieurs, un suivi à long terme de la glacière est proposé.**

INTRODUCTION

Les observations glaciologiques menées dès la fin du XIX^{e} siècle dans la chaîne alpine ont démontré un lien étroit entre les variations du bilan de masse des glaciers et les changements climatiques (e.g. Haeberli et al. 1999). Toutefois, peu d'informations existent sur l'évolution des remplissages de glace dans les cavités naturelles. Ces glacières présentent pourtant un réel intérêt scientifique pour l'étude du climat post-glaciaire dans des régions peu documentées (e.g. Perroux 2001, Luetscher et al. 2005).

Une approche théorique secondée d'observations en grottes a permis d'établir une relation directe entre le volume de glace souterraine et le climat hivernal extérieur (Luetscher et al. submitted). Il en ressort que les glacières observées de nos jours découlent essentiellement de processus d'accumulation actuels (ou récents) et ne devraient en principe pas être considérées comme fossiles. Dès lors, le bilan de masse annuel résulte d'une différence entre l'accumulation et l'ablation saisonnière de glace.

Une compilation de documents historiques permet un aperçu synthétique de l'évolution des glacières jurassiennes durant le XX^{e} siècle (Brulhart 2001, Luetscher et al. 2005). Celui-ci met en évidence un bilan de masse négatif pour l'ensemble des objets étudiés.

Photo: R. Wenger

Aspect particulier de la glacière de Monlési; accumulation de glace vue depuis le fond d'un moulin creusé par une arrivée d'eau.

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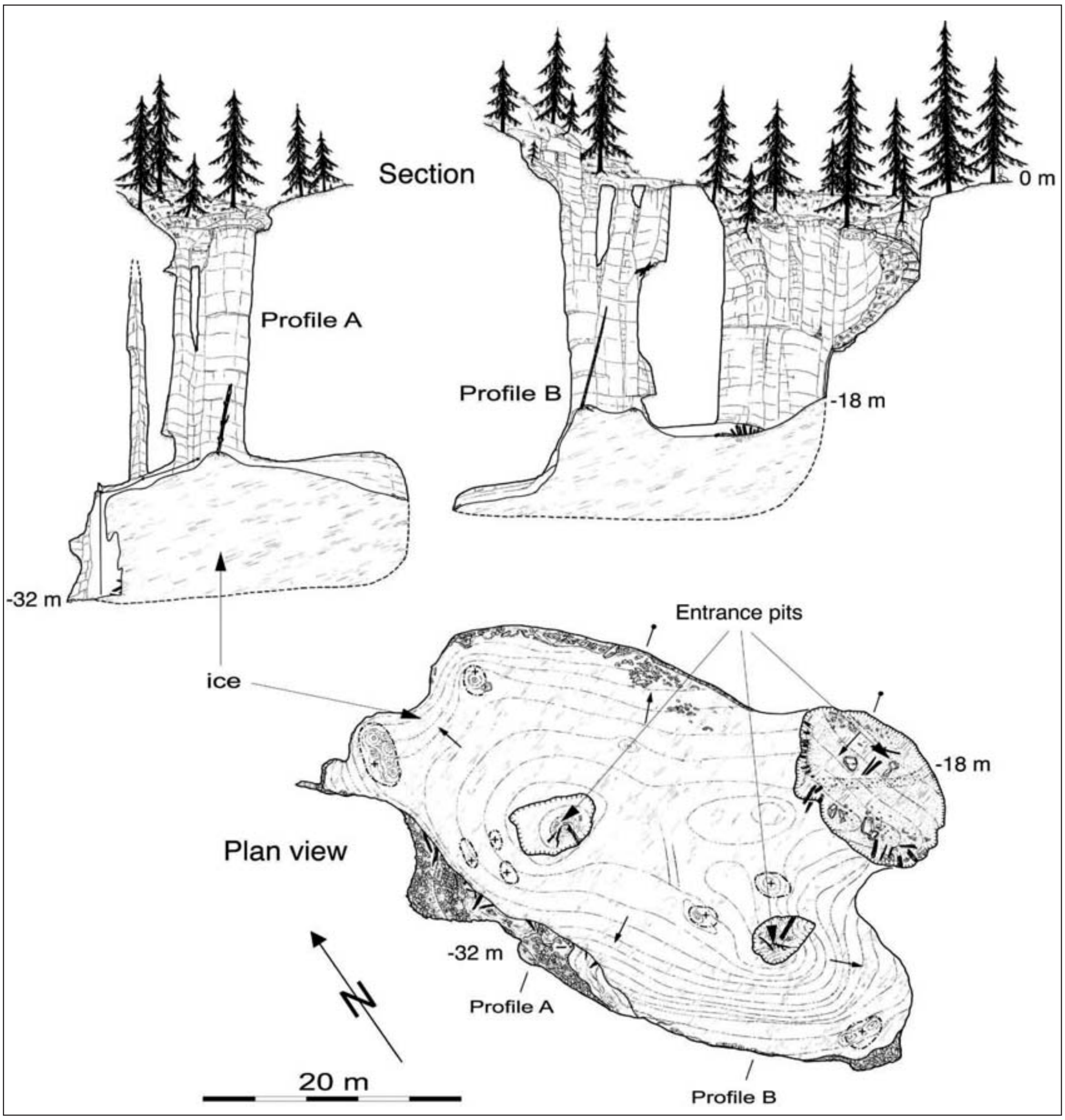


Fig. 1: Topographie de la glacière de Monlési (adaptée de Luetscher & Wenger 2002). L'emplacement des stations de mesures est symbolisé par une petite croix.

Particulièrement marquée depuis la fin des années 1980, cette tendance peut être attribuée à l'absence d'hivers rigoureux et à une diminution notable de l'enneigement à basse altitude. De fait, de nombreux spéléologues constatent avec désarroi une diminution de la glace souterraine dans la plupart des glacières de la chaîne jurassienne (e.g. Sesiano 1996). Peu documentées, ces observations qualitatives manquent cruellement de données mesurées permettant de quantifier les fluctuations volumétriques de ces remplissages. Cet article présente les premiers résultats issus d'observations entreprises à la glacière de Monlési entre 2001 et 2004.

MÉTHODOLOGIE

Située à 1130 m d'altitude (température moyenne extérieure d'environ 5°C), la glacière de Monlési (Boveresse, NE; fig. 1) possède le plus grand volume de glace hypogée reconnu dans la chaîne jurassienne. Celui-ci résulte essentiellement de glace de congélation provenant de la fonte des névés accumulés à la base des puits et de l'infiltration d'eau météorique circulant à la faveur de quelques fractures principales. Luetscher et Wenger (2002) ont estimé ce remplissage à quelque 6'000 m^3.

En vue de valider un schéma d'accumulation et de démontrer la tendance négative du bilan de masse de la glace, divers points de repères furent implantés dans la glacière dès le printemps 2001 (fig. 1).

La distance entre la glace et la roche fut mesurée à intervalles de temps variables aux 31 stations de mesures matérialisées au sein de la cavité par un repère de couleur sur la roche. Pour distinguer la présence éventuelle d'un fluage de ce volume de glace, 9 jalons supplémentaires furent implantés directement dans la glace. Leur emplacement relatif fut mesuré au moyen des techniques topométriques traditionnellement utilisées en spéléologie (Grossenbacher 1991). Celles-ci permettent dans le cas présent, d'atteindre une précision relative sur le positionnement de l'ordre de ± 5 cm en XYZ.

RÉSULTATS

Les mesures périodiques de hauteur de la glace sont présentées dans le tableau 1. Bien que n'étant pas levées à un pas de temps déterminé, ces mesures mettent clairement en évidence une accumulation saisonnière de glace, principalement active au printemps. Néanmoins, les valeurs mesurées augmentent au cours du temps, démontrant que la fonte annuelle est plus forte que l'accumulation printanière, induisant un bilan de masse négatif pendant la durée couverte par l'étude.

La figure 2 illustre cette tendance négative. Au cours de la période d'observation, la hauteur de la glace a décru de l'ordre de 10 cm par année au faîte du remplissage. Bien que cette valeur maximale ne soit pas parfaitement représentative de toutes les stations de mesures, il est possible d'évaluer pour la période 2001 à 2004 le déficit de glace entre 180 à 240 m^3 sur l'ensemble de la glacière (soit environ 160 à 220 tonnes de glace). Il est à relever que malgré un record de chaleur observé durant l'été 2003, aucune fonte supplémentaire n'est mise en évidence à ce moment là.

Le levé topographique des jalons implantés à même la glace met en évidence un léger fluage du remplissage en direction des parois de la glacière. Ces mesures sont étayées par de nombreuses observations de crevasses et de cassures de concrétions (par exemple colonnes de glace), illustrant au mieux ce déplacement. Celui-ci est attribué à la fusion du remplissage de glace au contact des parois de la cavité.

De même, un déplacement vertical de 20 à 30 cm est bien distinguable sur les repères implantés sur le flanc du remplissage, le long de la paroi de glace du puits terminal de la cavité (tabl. 2). Bien qu'une diminution de ce déplacement semble être mise en évidence en fonction de la profondeur, cette dernière observation reste fortement tributaire de l'incertitude des mesures. Dès lors l'interprétation de ce flux vertical peut être formulée comme étant la conséquence d'une fonte constante à la base du volume de glace. Cette fonte est attribuée au flux de chaleur issu du système karstique situé sous la cavité, estimé dans le cas présent à ~1 Wm^{-2}.

Bilan énergétique d'une visite de la glacière

En moyenne, chaque personne visitant la glacière dégage une énergie suffisante pour faire fondre l'équivalent d'un kilogramme de glace par heure. Si l'on estime la fréquentation de la glacière à environ 10 visites par semaine et que chacune de ces visites permet à quatre personnes de découvrir la cavité pendant une vingtaine de minutes, on en déduit que l'activité touristique à la glacière de Monlési induit la fonte de près de 740 kg de glace par année…

Bien que cette valeur puisse paraître considérable, ces visites n'ont contribué qu'à hauteur de 0,5 % au déficit de glace observé entre 2001 et 2004.

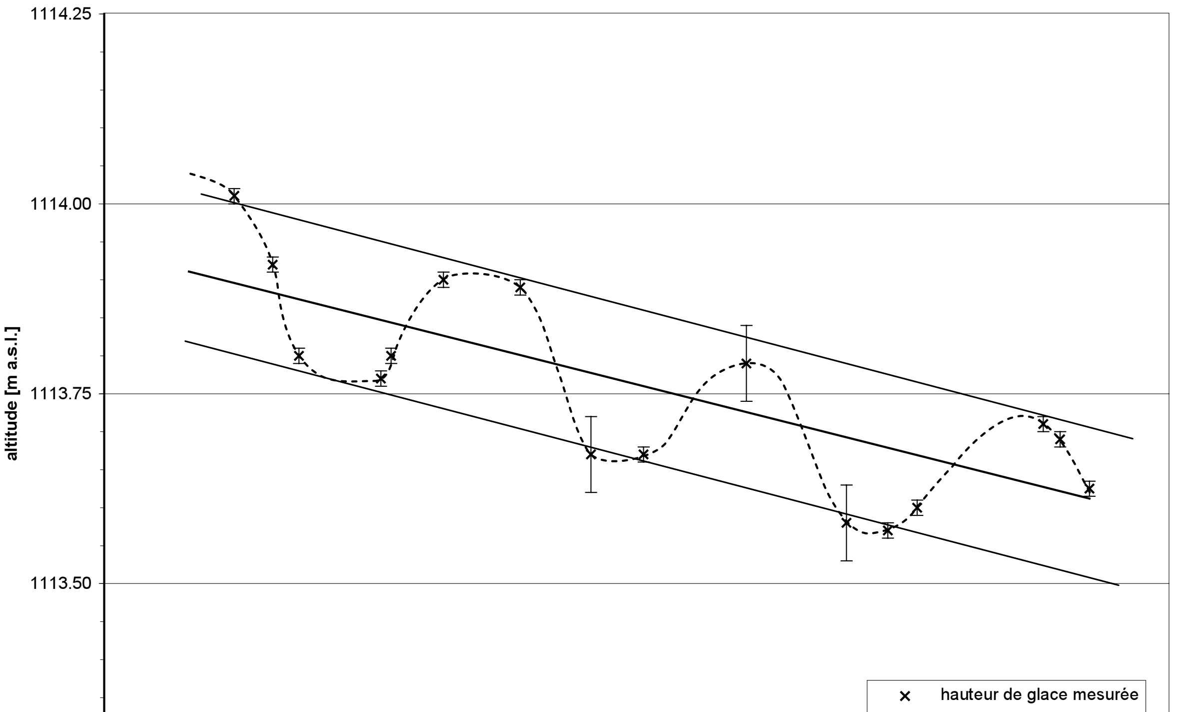
DISCUSSION

Les données acquises au cours de cette étude permettent de mieux conceptualiser les processus de dépôt et de fonte de glace au sein de la glacière de Monlési (fig. 3).

La différence de densité entre l'air souterrain et l'atmosphère extérieur induit d'importantes circulations d'air durant la période hivernale, favorisant ainsi le refroi-

Fig. 2: Évolution du niveau de glace à la glacière de Monlési entre 2001 et 2004. Bien que le bilan de masse esquisse une tendance négative, aucune fonte exceptionnelle n'est observable durant la canicule de l'été 2003.

▼



dissement de la glacière (Luetscher & Jeannin submitted, Luetscher & Jeannin 2002). Au printemps, grâce au froid accumulé dans la glacière, les premières infiltrations d'eau issues de la fonte des neiges constituent une nouvelle couche de glace à la surface du remplissage. Cette phase d'accumulation se poursuit tout au long du printemps, jusqu'à ce que la glacière atteigne l'équilibre thermique. Dès cet instant, toute nouvelle perturbation (eau, circulation d'air, visites,...) constituera une source d'énergie contribuant directement à la fonte de la glace. À cela s'ajoute la fonte constante de la base du remplissage provoquée par le flux de chaleur issu de système karstique sous-jacent. Pour permettre un équilibre du bilan de masse de la glacière de Monlési, une accumulation annuelle de plus de 200 m^3 de glace doit avoir lieu. Sans cela, le bilan de masse de la glace de Monlési restera inexorablement négatif.

CONCLUSION ET PERSPECTIVES

La surveillance d'un volume de glace souterraine a mis en évidence d'importantes fluctuations saisonnières. Suggéré par de nombreux auteurs, ces données quantitatives démontrent et confirment que la présence de ce type de remplissage résulte a priori de processus actuels. À la glacière de Monlési, la glace se renouvelle essentiellement au printemps, alors que la fonte à la base du remplissage est supposée avoir lieu tout au long de l'année. Cette fonte est attribuée au flux de chaleur issu du système karstique sous-jacent. Entre 2001 et 2004, la fonte annuelle étant néanmoins plus importante que la formation de nouvelle glace, un bilan de masse négatif peut être observé.

Dans un contexte climatique changeant, une surveillance à long terme des glacières est fortement souhaitable. En effet, si la communauté scientifique est unanime quant au réchauffement climatique en cours (IPCC 2001) peu de modèles sont disponibles pour

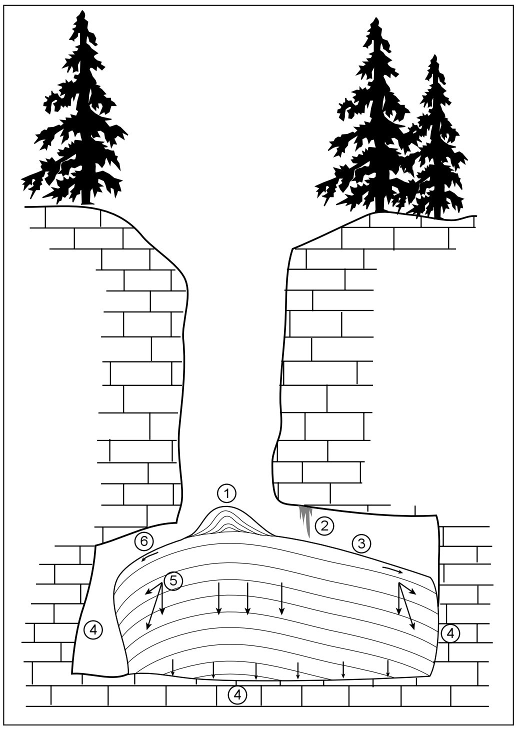


Fig. 3: Modèle de dépôt de glace au sein de la glacière de Monlési:
1 Accumulation hivernale de neige à la base des puits;
2 Infiltrations d'eau au printemps;
3 Cristallisation d'une nouvelle couche de glace;
4 Fonte du volume de glace au niveau des parois;
5 Fluage de la glace;
6 Fonte superficielle du volume de glace en période estivale.

permettre d'anticiper l'évolution des régimes climatiques hivernaux (températures et précipitations). Or les glacières dépendent essentiellement des conditions hivernales et les fluctuations volumétriques de glace souterraine constituent donc un marqueur intéressant de ces changements climatiques. En cela, le suivi de cette glace apporte un complément précieux à l'observation des glaciers alpins. Outre confirmer ces premières observations, un réseau de surveillance à long terme (à mettre en place) apportera les données nécessaires pour établir le lien entre les fluctuations climatiques extérieures et l'accumulation de glace printanière.

REMERCIEMENTS

Tous nos remerciements s'adressent aux nombreuses personnes ayant participé aux campagnes de mesures à l'origine de cet article. En particulier, nous remercions chaleureusement François Bourret pour sa contribution à la mise en place d'une surveillance à long terme de la glacière de Monlési. Les mesures effectuées durant cette étude s'inscrivent dans le cadre d'un projet soutenu par le Fonds national suisse pour la recherche scientifique (SNF 200020-103538/1).

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n° station	Y / X / Z [m] / [m] / [m]	24.04 2001 [cm]	28.06 2001 [cm]	20.08 2001 [cm]	28.08 2001 [cm]	01.09 2001 [cm]	15.01 2002 [cm]	29.02 2002 [cm]	11.04 2001 [cm]	25.07 2002 [cm]	10.01 2003 [cm]	13.02 2003 [cm]	11.12 2003 [cm]	20.01 2004 [cm]	11.07 2004 [cm]
17.5	534910.89 / 198923.54 / 1115.72	165	168	n.d.	177	n.d.	198	196	184	186	n.d.	n.d.	218	n.d.	204
18.1	534913.19 / 198924.58 / 1115.82	n.d.	175	184	186	196	199	196	186	187	209	n.d.	219	216	205
S1_11	534930.42 / 198910.26 / 1114.82	n.d.	n.d.	n.d.	n.d.	n.d.	93	92	82	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
S1_12	534928.80 / 198911.08 / 1114.83	n.d.	n.d.	n.d.	n.d.	n.d.	99	95.3	88	n.d.	103	109	n.d.	n.d.	112
S1_13	534928.90 / 198913.57 / 1114.95	n.d.	n.d.	n.d.	n.d.	n.d.	107	107	96	90	101	114	123	n.d.	118.8
S1_14	534928.94 / 198915.55 / 1115.08	n.d.	n.d.	n.d.	n.d.	n.d.	134	135	130	127	136	144	152	147	149
S1_15	534928.04 / 198917.38 / 1114.76	n.d.	n.d.	n.d.	n.d.	n.d.	163	164	160	160	167	170	175	n.d.	172.1
S1_16	534927.45 / 198919.32 / 1115.32	n.d.	n.d.	n.d.	n.d.	n.d.	213	212	207	205	208	217	223	n.d.	218.5
S1_17	534926.05 / 198920.82 / 1115.26	n.d.	n.d.	n.d.	n.d.	n.d.	224	223	222	221	224	234	239	n.d.	230
S1_18	534924.52 / 198922.15 / 1115.26	n.d.	n.d.	n.d.	n.d.	n.d.	225	228	225	224	221	240	241	n.d.	235.8
S1_19	534923.15 / 198923.49 / 1115.18	n.d.	n.d.	n.d.	n.d.	n.d.	214	215	211	213	215	223	227	n.d.	228
S1_20	534922.12 / 198924.95 / 1114.46	n.d.	n.d.	n.d.	n.d.	n.d.	170	170.5	168	167	170	177	184	n.d.	184.5
S1_21	534921.00 / 198926.42 / 1114.33	n.d.	n.d.	n.d.	n.d.	n.d.	152	154	154	151	157	162	168	n.d.	167.2
S1_22	534919.41 / 198927.65 / 1114.33	n.d.	n.d.	n.d.	n.d.	n.d.	127	129	124	122	124	137	141.5	n.d.	140.2
S1_23	534917.79 / 198928.92 / 1114.43	n.d.	n.d.	n.d.	n.d.	n.d.	138	134	133	134	140	105.2	151	150	149.5
S2_14	534914.55 / 198917.80 / 1114.04	n.d.	n.d.	n.d.	n.d.	n.d.	60	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
S2_15	534914.05 / 198916.30 / 1113.57	n.d.	n.d.	n.d.	n.d.	n.d.	44	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
S2_16	534913.17 / 198914.56 / 1113.38	n.d.	n.d.	n.d.	n.d.	n.d.	40	40	n.d.	30	n.d.	n.d.	n.d.	n.d.	n.d.
S2_17	534912.68 / 198912.69 / 1113.28	n.d.	n.d.	n.d.	n.d.	n.d.	63	62	57	58	n.d.	n.d.	n.d.	n.d.	n.d.
S2_18	534911.25 / 198911.29 / 1113.15	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	55	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.0	534930.65 / 198913.37 / 1114.89	n.d.	93	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.1	534931.22 / 198919.88 / 1113.55	n.d.	122	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.2	534926.79 / 198928.51 / 1112.51	n.d.	170	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.3	534920.24 / 198935.41 / 1111.95	n.d.	134	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.4	534914.87 / 198937.50 / 1111.87	n.d.	168	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.5	534911.75 / 198935.44 / 1112.03	n.d.	120	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.6	534908.04 / 198930.06 / 1113.40	71	77	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
6.7	534910.09 / 198926.99 / 1114.18	n.d.	53	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
8.10	534912.58 / 198918.70 / 1114.37	n.d.	62	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

Tab. 1: Distances mesurées entre les stations de références et le niveau de glace.

2a

Date	N°	Y [m]	X [m]	Z [m]	DY [m]	DX [m]	DZ [m]
11.02.2004	1	534927.45	198915.23	1114.89			
11.02.2002	2	534926.02	198918.88	1113.51	0.20	0.00	0.22
11.07.2004	2	534926.22	198918.88	1113.29			
11.02.2002	3	534919.60	198917.27	1113.77			
11.02.2002	4	534922.04	198924.85	1112.87	0.09	0.06	0.16
11.07.2004	4	534922.13	198924.91	1112.71			
11.02.2002	5	534917.38	198928.69	1113.31	0.24	0.15	0.49
11.07.2004	5	534917.62	198928.84	1112.82			
11.02.2002	6	534908.77	198927.06	1113.56			

2b

Date	N°	Distance [m]	Azimut [G]	Pente [G]	DH [G]
08.10.2001	clou 1	3.41	61	22	-0.28
23.03.2004	clou 1	3.34	61	18	
12.07.2004	clou 1	3.33	61	17	
08.10.2001	clou 2	3.28	61	11	-0.25
23.03.2004	clou 2	3.42	61	7	
12.04.2004	clou 2	3.42	61	6	
08.10.2001	clou 3	3.26	61	0	-0.21
23.03.2004	clou 3	3.37	61	-3	
12.04.2004	clou 3	3.37	61	-4	

Tab. 2 : Fluage de la glace mesuré aux points de repères implantés dans la glacière de Monlési.
2a: stations implantées sur la surface de la glace.
2b: stations implantées sur la coupe verticale de la glace.

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Energy fluxes in an ice cave of sporadic permafrost in the Swiss Jura mountains – concept and first observational results

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ABSTRACT: Low-altitude perennial ice occurrences are observed in ice caves of the Jura mountains. Resulting from a fragile climatological equilibrium, the presence of ice is subject to significant variations. The paper aims at presenting first observations on distribution in time and space of heat transfer processes taking place between air, ice and rock, and their influence on ice formation and conservation of ice. Correlation with rock and air temperatures allows to present a simplified model of heat transfer processes with the surrounding limestone. In a first approach, only conductive heat exchanges are considered. In order to calibrate the model, observations are carried out about the ice mass of "La Glacière de Monlési" (Boveresse, Swiss Jura) at an altitude of 1135 m. Continuous temperature measurements permit a better definition of its thermodynamical characteristics and document the role played by water infiltration and convective heat transfers by air circulation around the ice filling.

1 INTRODUCTION

1.1 *Previous studies*

Known for a long time by local people, ice caves have been of interest to European naturalists since the second part of the XVIIIth century. Various hypotheses, more or less credible, were established on the genesis of such fillings (e.g. Kyrle 1923), among them a first real scientific work can be mentioned: Thury (1861) made a detailed description of three ice caves in the Jura Mountains at different seasons. This author first displayed the importance of subsurface climate for the perennial preservation of ice in a cave.

Throughout the last century several authors initiated larger studies on some particular ice caves. Long-term measurements were started in Scarisoara Ice Cave/Romania (e.g. Silvestru 1999, Racovita 1994, Viehmann 1976) and several meteorological measurements were carried out in the Dachstein massif/Austria (e.g. Mais 1999, Pavuza & Mais 1999) and Dobsinska Ice Cave/Slovakia (e.g. Lalkoviè 1995).

These numerous studies brought a lot of field observations and permitted a better understanding of different phenomena occurring in ice caves. However, only few of the presented ideas concerning the formation of cave ice are based on the physics of heat transfers in such cavities.

Notions of static and dynamic caves were already proposed by Thury (1861), but a real physical approach of subsurface climate was stimulated by the work of Trombe (1952). During the last fifty years, several authors completed important studies on this particular subject (e.g. Cigna 1968, Andrieux 1969, Eraso 1978). Encouraged by application on prehistorical sites, the research on cave temperatures and air circulation considerably improved the data acquisition and their interpretation (e.g. Andrieux 1983).

Badino (1995) and Lismonde (2002) proposed the first synthesis handbooks on cave climatology. However, the special problem of the physics of ice caves was still rather marginally considered. Our research aims at filling the gap by applying heat transfer and fluid dynamics modelling to the study of ice caves. The present note describes the first ideas developed to this purpose. Ice caves of the Jura Mountains are used for observations.

1.2 *Definition and nomenclature*

Luetscher & Jeannin (2001) outlined the considerable confusion existing today in the nomenclature about ice caves. In order to facilitate the understanding of the following text, ice caves are defined as:

"Natural karstic cavities presenting a perennial ice filling."

Such ice caves have been recognized for a long time as special features of perennially frozen ground, sometimes occurring at very low altitudes (Haeberli 1978).

The wide definition used here is independent of any type of morphological cave description and different ice origins.

2 SITE DESCRIPTION

2.1 *Ice caves in the Jura Mountains*

The Jura Mountains form a ~400 km wide arc between the Savoie (France) in the South and the Black Forest (Germany) in the North. Their total surface is estimated at about 14,000 km^2, and their highest peaks reach about 1700 m a.s.l. Mainly composed of karstic terrains, this massif includes over 10,000 known natural cavities. The altitude of the 0°C-isotherm of mean annual air temperature, is estimated in this region at approximately 2250 m a.s.l. Far below this level, the Jura Mountains include about 50 ice caves, located between 1000 and 1500 m a.s.l., with an ice filling volume varying between a few cubic meters up to more than 5000 m^3.

2.2 *The Monlési ice cave*

Located in the Swiss Jura Mountains (Boveresse/Ne, 6°35′4″/46°56′18″, 1135 m), the "Glacière de Monlési" is the largest Ice Cave in the Jura. First descriptions of its important ice filling were provided by Browne (1865), while further studies were initiated by Stettler & Monard (Stettler & Monard 1960, Stettler 1971).

Three entrance shafts allow to reach a large room of approximately 25 m × 45 m at a depth of 20 m below surface. Partially filled with an important mass of congelation ice, the lowest point of the cave can be reached at −33 m through a large "rimaye" located on the south-western part of the "glacier". Luetscher & Wenger (2002) estimated the total ice volume to be about 6000 m^3.

Unfortunately, the thickness of the ice mass cannot be measured precisely. The presence of numerous cryoclastic sediments (e.g. Pancza 1992) and wood inclusions in the ice make the use of traditional geophysical methods (e.g. GPR) difficult. In the same way, steam drilling could not yet reach the bottom of the total ice mass.

Up to now only few descriptions were prepared on the ice itself. The access provided by the "rimaye" enables observation of a 15 m ice thickness. Stratification within the ice seems to indicate a maximum age of 100 years for the lower levels of the observed sequence. This age is confirmed so far by the dating of a tile included in the ice mass. Due to the young age of the material, dendrochronological correlation and dating could not be carried out on the numerous wood inclusions.

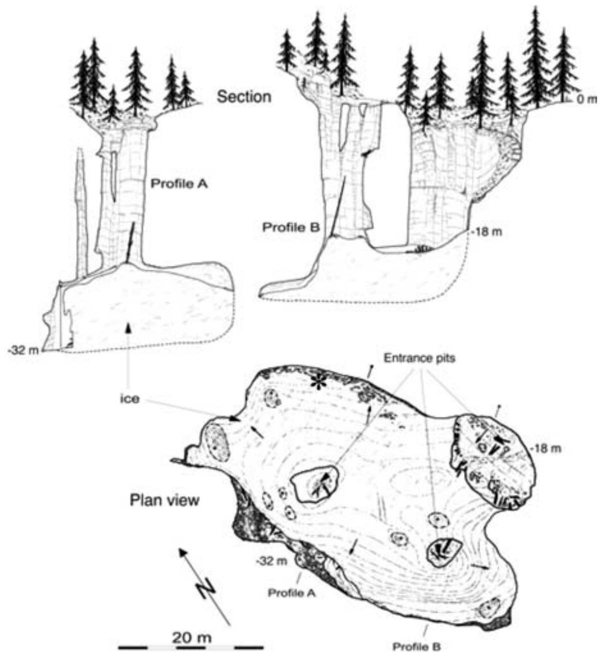

Figure 1. Plan view and vertical cross sections of the "Glacière de Monlési" – adapted from Luetscher & Wenger 2002. The position of the minilogger is mentioned by an *.

An ice core sample could furnish further information on the age of the ice. At the same time, isotopic analysis may provide additional information on climatic conditions in the Jura Mountains.

3 ENERGY BALANCE AND HEAT EXCHANGE ESTIMATION

3.1 *Definition of the system*

An ice cave can be described as a system controlled by the surface climatological conditions and those prevailing in the surrounding karstic massif.

This system is defined as an ice volume filling a vertical shaft. The ice column is in contact with the surrounding limestone on its lower part and along the vertical walls. The upper ice surface is in contact with the cavity air (fig. 2). As a first approach, necessarily a bit simplistic, we will only consider a two-dimensional system.

3.1.1 *Basic assumptions*

The filling is expected to be essentially constituted of congelation ice (e.g. Shumskii 1964). It is assumed that the cave acts as a cold air trap, i.e. air circulations due to the possible presence of several entrances are neglected in this first approach.

In order to understand the thermal processes related to the thermodynamic behaviour of ice caves, energy

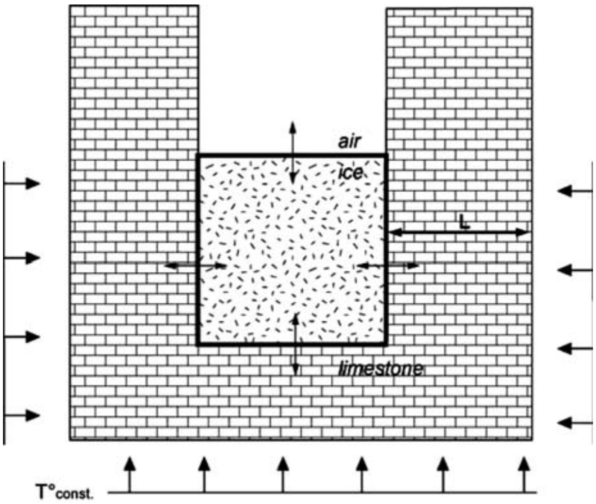

Figure 2. Schematic representation of the studied system. Energy transfers are represented by the different arrows. As a first assumption, the model is considered to be a cold air trap. L: distance between the cave and a constant rock temperature.

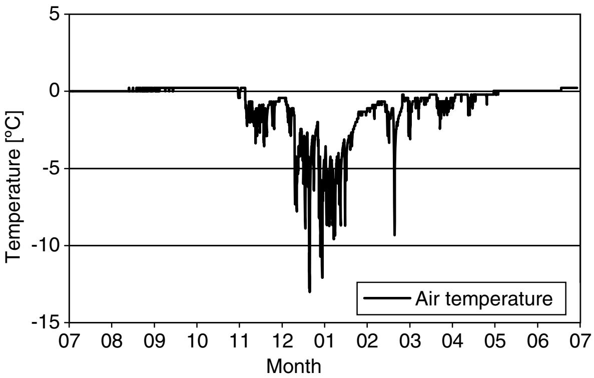

Figure 3. Air temperatures acquired during a one year cycle in the Monlési ice cave present a typical cold air trap signature.

exchanges between the filling and its surroundings must be determined.

3.1.2 *Boundary conditions*

Air temperatures in the cave constitute the upper boundary condition of the defined system. Due to melting, these temperatures cannot rise above 0°C and the ice surface (i.e. cave air) has a well defined thermal maximum. As air exchanges within the cave are restricted to winter time, minima are defined by air temperature during cold periods. Figure 3 presents a typical temperature record acquired in an ice cave which acts as a cold-air trap.

Lower boundary conditions are characterised by a constant rock temperature at a certain depth. It can be easily demonstrated that, due to the presence of a high water flux in karstic terrain, geothermal heat flux is negligible and rock temperatures are controlled by water and air temperatures (e.g. Badino 1995, Jeannin et al. 1997). The distance to the cave at which this fixed temperature boundary has to be set is not known so far. Depending on the fractures surrounding the cave, it may range between some meters to tens of meters.

3.2 *Estimation and discussion of the respective heat fluxes*

3.2.1 *Ground heat flux*

In order to estimate the role of ground heat fluxes, steady-state, one-dimensional, conductive heat transfers are considered. To this purpose, the surrounding limestone is expected to be a homogeneous body ($\lambda = 2.2\ \mathrm{Wm^{-1}K^{-1}}$) without any water infiltration. Heat fluxes due to conductive energy exchanges through the surrounding limestone can be expressed with the following equation:

$$_g = S\lambda_r \frac{T_m - T}{L} \tag{1}$$

where $_g$ = ground heat flux [W]; S = exchange surface [$\mathrm{m^2}$]; λ_r = thermal conductivity of limestone [$\mathrm{WK^{-1}m^{-1}}$]; T_m = rock temperature at the fixed temperature boundary; T = temperature at ice-rock interface; L = distance [m].

Assuming a (constant) local temperature boundary of 5°C (the ice-rock interface is assumed to be at melting point, i.e. 0°C) it is possible to estimate a plausible distance for "L" ranging between 3–20 m. Below this value, ice melt would probably be too important for a long time conservation of the filling. On the other hand, higher values than 20 m would mean that conductive heat exchanges through the surrounding limestone hardly affect the ice and would enable ice to grow up close to the top of the shaft.

Sensible heat fluxes characterise the heat storage in the ice and surrounding limestone as related to their temperature changes. The cooling of the cave (air, but especially rockwalls and ice) during the winter season will therefore contribute to the total inertia of the system.

3.2.2 *Fluxes at ice-air interface*

Due to the geometry of the cave entrance and the vegetation around it, direct effects from radiative heat fluxes can often be neglected. Hoelzle et al. (1993) already provided evidence for the possible presence of low-altitude ice occurrences under similar conditions.

More important could be the role of latent heat fluxes induced by air circulation which will largely favour phase changes of water (sublimation, evaporation and condensation). Besides their influence on heat exchanges, condensation processes could possibly play a non-negligible role for the ice origin (Bock 1913).

Convective heat exchange with surface air have not been approached yet. Their importance is, however, essential in the final energy balance of the system and is at the origin of the cold air present in the cave.

3.2.3 *Freezing and melting fluxes*

Freezing, essentially affecting water infiltrated from the surface or more rarely meltwater from the ice, is usually restricted to the winter season. As freezing is an exothermal process, the released energy contributes to the warming of the cave. Negative temperatures will, therefore, be rapidly equilibrated and very low temperatures will not persist for long periods as far as infiltration provides enough water. Convective exchanges with outside air induces warm air flowing out and being replaced by colder air. This contributes to the cooling of the massif (Lismonde 2002).

If the ice filling is assumed to be in equilibrium, melting (mostly during summer) must compensate the total ice produced in the cave during winter. Plausible values for the yearly ice deposit are in an order of 100–500 kg/m^2. The energy necessary to melt such a volume of ice corresponds to a heat flux varying between 1–5 W/m^2 over the entire year. The less the ice volume is, the more the latent heat will constitute the essential factor influencing the thermal inertia of the cave.

Water infiltration is necessary for the formation and presence of ice but, also for the supply of an important part of the energy used to melt the ice filling. With annual precipitations of some 1500 mm and a temperature difference of 5°C between the in- and output of the water-infiltration system, this energy supply can be estimated at about 1 W/m^2.

Snow accumulation constitutes an important cold reservoir and largely contributes to the total inertia of the filling.

3.2.4 *Human impact*

Human impact can under certain circumstances play an important role on ice conservation/disappearance. Besides the energy supply resulting from each visitor, visits influence the natural state of the cave by enhancing the exchange of air (turbulences) with the surface and, hence, by disturbing a rather static environment.

Ice exploitation was frequent in certain caves during the 19th century and, could also have considerably disturbed the equilibrium permitting the conservation of ice. However, some ice caves seem to have completely recovered.

3.3 *Energy balance of the proposed model*

The energy balance of the defined system can now be expressed as following:

$$\int_{t=0}^{1\,year} (S + M + H + LE + R + A) = 0 \qquad (2)$$

Where S = ground heat flux, M = heat flux due to surface melt or freezing water, H = sensible heat flux, LE = latent heat flux, R = radiative heat flux, A = anthropogenic heat flux.

Two major processes control this energy balance, i.e. the presence of ice: water infiltration and air exchanges.

The importance for the genesis and conservation of the ice filling from water infiltration is evident. Each cave must have its own infiltration optimum depending on geometric characteristics of the conduits with their spatial frequency and fluxes involved.

Too little infiltration prevents ice to form because of the absence of water, too much infiltration brings too much heat into the cave and ice melts. A cavity connected to fissures and conduits leading deeper into the karst will "breath" a lot as a function depending on inside/outside meteorology. High heat fluxes are thus transferred through the cave and ice cannot occur at low elevation. A cave closed at its bottom will act as a cold trap and represents good conditions for ice to stay.

The penetration depth of the fractures near the cave walls is important: depending on the length scale of the air and water transportation processes involved, the system boundary might correspond to a fixed temperature bringing a high heat flux into the cave.

In order to better quantify the importance of each parameter, field observations are carried out in several ice caves of the Jura Mountains with one of them, the Monlési Ice Cave, being completely instrumented. A better understanding of ice conservation processes should thereby become possible.

4 FIELD OBSERVATIONS

4.1 *First temperature records*

As a first step of the observational programme, air temperatures were acquired on an annual cycle with "UTL-1 Miniloggers" developed by the University of Bern/CH (Hoelzle et al. 1999). A mean annual air temperature of −0.9°C could be established during the first annual cycle recorded (07.01–06.02) in the Monlési ice cave (Fig. 3). The minilogger was located in the rimaye of the northern part of the cave.

The collected data document the importance of the thermal inertia induced by the melting of the ice filling. The presence of ice prevents temperature to rise above 0°C during the whole summer season.

The absence of significant water infiltrations during winter induced an important cooling of the subsurface air.

Boreholes temperatures in the ice are measured since march 2002 with platine thermistors (Pt 100) at different depths in the ice filling (0.1 m; −1 m;

−2,5 m; −4 m; −5 m; −8 m). Due to many impurities of the ice (stones and wood), measurements at depths below 8 m are not yet possible. As a consequence, the recorded temperatures are still influenced by seasonal variations. Long-term measurements are expected to provide a better understanding on the inertia resulting from sensible heat flux. First values acquired during March 2002 already indicate temperatures close to the melting point.

In the same way, rock temperature are recorded at different depths in small drill holes. These measurements improve our knowledge on water circulations affecting the surroundings of the cave and help to better define the importance of the rock heat flux.

4.2 *Water infiltration*

More than ten temporary infiltration points (from less than 0.1 l/min up to more than 10 l/min) have already been recognised in the cave. The importance of each of them must be quantified in order to determine the volume of ice produced and the melting occasioned by the energy supply. Frequency and flow will play a major role in the final mass and energy balance of the filling.

5 CONCLUSION AND PERSPECTIVES

Consideration of conductive heat transfer processes in the ice filling of an ice cave helps with the design of an observational programme. The strong inertia provided by the presence of ice points to marked effects of auto-regulation of the underground air temperature by storage of "cold" in ice and rockwalls.

Comparison with long-term field measurements indicates the *a priori* importance of water infiltration as related to conductive heat transfers as well as for the final mass balance of the ice. An identification of the respective role of each parameter should allow a better knowledge of heat exchanges in an ice cave. The understanding of the climatological conditions determining ice preservation should be considerably improved by the combination of field observations and numerical modelling.

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Une année d'enregistrement de températures à la glacière de Monlési (NE)

Ein Jahr Temperaturmessungen in der Glacière de Monlési (NE)

Les températures de l'air mesurées pendant une année dans la glacière de Monlési permettent de mieux comprendre le régime climatique de cette cavité. La comparaison entre les températures de surface et souterraines met en évidence le fonctionnement en piège à air froid alors qu'un enregistrement sur une année esquisse le comportement thermique de la glacière. Cette étude préliminaire livre les bases des travaux nécessaires à la compréhension des processus de formation et de conservation de la glace dans ces cavités situées bien en dessous des limites habituelles du permafrost.

Temperaturmessungen während eines Jahres in der Eishöhle « Glacière de Monlési » erlaubten, das Klima dieser Höhle besser zu verstehen. Der Vergleich zwischen den Temperaturen in und ausserhalb der Höhle zeigt das Funktionieren als Kaltluftfalle, während die Dauerregistrierung das thermische Verhalten des Eishohlraumes beschreibt . Diese Vorstudie bildet eine Grundlage für weitere vorgesehene Arbeiten mit dem Ziel, die Prozesse der Bildung und Erhaltung von Eis in Höhlen, die unterhalb der üblichen Grenzen des Permafrosts liegen, zu untersuchen.

Introduction

Sur quelque 10'000 cavités recensées dans la chaîne jurassienne franco-suisse (Blant 2001), seule une cinquantaine présente un remplissage permanent de glace. L'existence de ces « glacières » repose sur un fragile équilibre entre la morphologie particulière de la cavité et des conditions climatiques favorables.

Afin de mieux comprendre les processus à l'origine de la genèse et de la conservation de ce type de remplissage, une étude approfondie du microclimat souterrain s'avère indispensable. Cette note présente les premières observations obtenues après une année de mesures des températures de l'air dans la « Glacière de Monlési » (Boveresse / NE). Pour une description détaillée de la cavité, le lecteur se référera à la bibliographie existante (Luetscher & Wenger 2002 ; Gigon 1976 ; Stettler 1971).

Contexte climatologique

La glacière de Monlési, située à quelque 1100 m d'altitude dans le Jura neuchâtelois, bénéficie d'un climat fortement influencé par la côte atlantique. Les précipitations dans cette région sont abondantes (1500 mm / an) et la température moyenne annuelle à cette altitude est estimée à 5.5°C. Un pergélisol ne peut dès lors exister que dans quelques rares situations bien précises, et dépend de nombreux paramètres topo-climatiques.

Entourée d'un bosquet d'arbres, la glacière de Monlési semble *a priori* à l'abri de fortes rafales de vent et,

Einleitung

Von den gut 10'000 (Blant 2001) beschriebenen Höhlen in der französischen und schweizerischen Jurakette weisen nur etwa deren 50 eine permanente Eisfüllung auf. Diese unterirdischen « Gletscher » beruhen auf einem fragilen Gleichgewicht zwischen der spezifischen Morphologie der Höhle und günstigen klimatischen Bedingungen.

Um die Prozesse besser zu verstehen, die zur Eisbildung und -erhaltung führen, ist ein gründliches Studium des unterirdischen Mikroklimas unerlässlich. Die vorliegende Arbeit präsentiert die ersten Messungen der Lufttemperaturen in der Eishöhle « Glacière de Monlési » (Boveresse/NE). Für die Beschreibung der Höhle siehe Luetscher & Wenger (2002), Gigon (1976) und Stettler (1971).

Das klimatologische Umfeld

Die Glacière de Monlési liegt auf gut 1100 m Höhe im Neuenburger Jura in einem Klima, das von der atlantischen Küste beeinflusst wird. In dieser Region gibt es ausgiebige Niederschläge (1500 mm/Jahr) und die Jahresmitteltemperatur beträgt 5,5°C. Demzufolge kann hier Permafrost nur unter ganz speziellen, seltenen Bedingungen auftreten und muss von vielen topographisch/klimatischen Parametern abhängen.

Umgeben von einem Wäldchen ist die Glacière de Monlési vor Wind geschützt, und trotz den grossen Ein-

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Übersetzung
Hans Stünzi

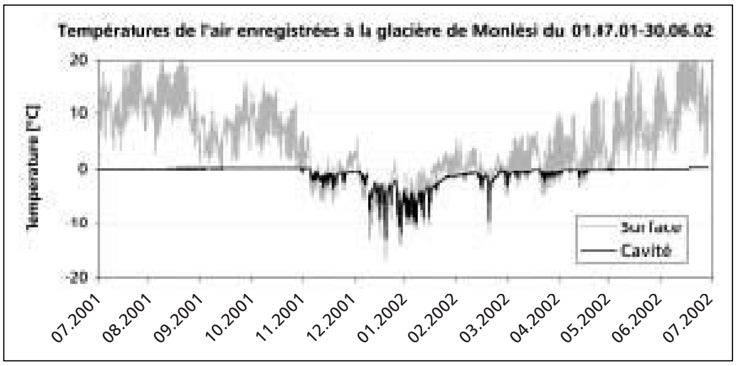

Fig. 1 : La chronique annuelle des enregistrements de température à la glacière de Monlési indique une bonne corrélation entre les fluctuations hivernales de la température extérieure et la température mesurée à l'intérieur de la cavité.

Abb. 1 : Die Temperaturaufzeichnung in der Glacière de Monlési während einem Jahr zeigt für den Winter eine gute Korrelation mit den Aussentemperaturen.

bien que possédant de larges entrées, la couverture végétale dense permet de fortement atténuer l'influence du rayonnement solaire. Les trois puits d'entrée favorisent en outre l'accumulation de neige à leur base durant la période hivernale.

Méthodes de mesure

Cette étude préliminaire vise en premier lieu une mise en évidence des caractères climatiques généraux de la glacière. Dans ce but, l'acquisition s'est concentrée sur deux points de mesures, l'un en surface et l'autre à l'intérieur de la cavité.

L'utilisation de Minilogger UTL-1, développés à l'université de Berne, s'est avérée particulièrement efficace pour cette première phase d'acquisition de données. D'une manipulation aisée et d'un prix accessible, ces instruments permettent un enregistrement de la température à un pas de temps défini pour une résolution de 0.27°C dans la plage de mesure concernée (-30/+40 °C). Bien que relativement élevée, cette valeur ne porte aucun préjudice aux observations effectuées dans notre cas.

L'étalonnage systématique de chaque instrument est indispensable. Nous l'avons fait sur un seul point (0°C), correspondant à de la glace fondante.

Résultats et interprétation

Une rapide comparaison entre les températures de surface et les températures souterraines (fig.1) révèle une corrélation presque parfaite pour des valeurs extérieures négatives. Lorsque ces dernières sont positives, la température hypogée marque toutefois un important palier aux alentours de 0°C, dû à l'influence de la chaleur latente de fusion de la glace.

La fluctuation des températures est limitée à une période précise de l'année (novembre à avril), et confirme le fonctionnement en piège à air froid de la cavité. Durant cette période (période ouverte), l'échange d'air avec l'atmosphère extérieure est caractérisé par un signal extrêmement nerveux, répondant aux variations de la température extérieure. Une périodicité journalière peut être observée sur ce signal, correspondant à l'alternance entre le jour et la nuit. Dans la glacière, cette période est caractérisée par de fortes circulations d'air (plusieurs m3/s) permettant un échange important entre la cavité et l'extérieur.

La figure 2 indique que le refroidissement occasionné par une chute brutale de la température extérieure

[1] **Indice de gel** : Mesure combinée durée-amplitude des températures inférieures à 0°C durant une saison de gel déterminée. L'indice est calculé en additionnant le nombre de degrés-jours inférieurs à 0°C et en soustrayant du total le nombre de degrés-jours supérieurs à 0°C au cours de la même période.

[1] **Frostindex** : Ein kombiniertes Mass aus der Dauer und Stärke von Temperaturen unter Null, die im Laufe einer Frostperiode auftreten. Er berechnet sich aus der Akkumulierung von Gradtagen unter 0°C und durch Subtraktion hiervon der Zahl von Gradtagen über 0°C während derselben Periode.

gängen dürfte die dichte Vegetationsdecke den Einfluss der Sonneneinstrahlung abschwächen. Im weiteren erlauben die drei Eingangsschächte eine Anhäufung von Schnee während des Winters.

Messmethode

Diese Vorstudie soll in erster Linie den allgemeinen klimatischen Charakter dieser Eishöhle hervorheben. Zu diesem Ziel hat sich die Datensammlung auf zwei Messpunkte konzentriert, je einen an der Oberfläche und in der Höhle.

Die Verwendung der Minilogger UTL-1, die an der Uni Bern entwickelt wurden, erwies sich in dieser ersten Phase der Datenakquisition speziell effizient. Die Instrumente sind einfach zu bedienen, haben einen erschwinglichen Preis und erlauben die Aufzeichnung mit wählbarem Messintervall. Die Temperaturauflösung beträgt 0.27°C bei einem Messbereich von -30° bis +40°C. Diese relativ mässige Auflösung genügte vollauf für eine Vorstudie. Alle Geräte wurden bei 0°C (in schmelzendem Eis) kalibriert.

Resultate und Interpretation

Schon ein schneller Vergleich zwischen den ober- und unterirdischen Temperaturen (Abbildung 1) zeigt eine gute Korrelation bei negativen Aussentemperaturen. Wenn die letzteren positiv sind, bleibt die Höhlentemperatur um 0°C, was der latenten Schmelzwärme des Eises zugeschrieben werden kann.

Nach der konstanten Sommertemperatur folgen Temperaturschwankungen in der Zeit von November bis April. Dies bestätigt die Funktion der Höhle als Kaltluftfalle. Während dieser Zeitspanne mit Luftaustausch mit der Aussenluft (« offene Periode ») ist das Temperatursignal extrem nervös und folgt den äusseren Temperaturschwankungen. Die Höhlentemperatur zeigt eine Tagesperiodizität im Tag/Nacht-Rhythmus. Zu dieser Jahreszeit erlauben starke Luftzirkulationen (mehrere m^3/s) einen signifikanten Luftaustausch mit der Aussenwelt.

Abbildung 2 zeigt, dass ein äusserer, brutaler Temperatursturz nur eine relativ kurz andauernde Abkühlung in der Höhle zur Folge hat. Dieses Phänomen kann teilweise durch Eisbildung erklärt werden, denn dieser Prozess setzt Wärme frei und beschleunigt die Einstellung des Gleichgewichts zwischen Eis und umgebendem Fels. Die häufigen Temperaturschwankungen in dieser Jahreszeit bewirken denn auch eine ausgedehnte Frostsprengung im umgebenden Gestein (Pancza 1992).

Die Beobachtungen im Winter 2001/02 zeigen eine bedeutende Abkühlung der Höhle vom 8. Dezember bis 16. Februar. Diese Abkühlung wird der Abwesenheit von einfliessendem Wasser zugeschrieben (keine wesentlichen Niederschläge) ; somit wird der Höhle keine Energie zugeführt.

Die Abkühlung der Höhle während eines Jahres kann aus dem Integral der Temperaturkurve während eines Winter-Zyklus abgeschätzt werden. Der damit erhaltene Frostindex[1] erlaubt einen guten Vergleich der Abkühlung, die von Jahr zu Jahr auf eine Struktur wirkt. Die Messungen im Winter 2001/2002 ergaben einen Frostindex von 300°C•Tag. Der Vergleich dieses Wertes mit semble de courte durée à l'intérieur de la cavité. Ce phénomène peut être partiellement expliqué par la formation de glace dans la grotte. Ce processus libère en effet de la chaleur et accélère la mise à l'équilibre des températures de l'air et de son encaissant. Ces fluctuations de la température, nombreuses au cours d'une saison, favorisent une importante gélifraction de la roche encaissante (Pancza 1992).

La chronique observée durant l'hiver 2001/2002 met clairement en évidence un important refroidissement de la cavité entre le 08 décembre et le 16 février. Ce fort refroidissement est attribué à l'absence d'infiltration d'eau durant cette période (pas ou peu de précipitations observées sur le site de Monlési), empêchant tout apport d'énergie dans la cavité.

Le refroidissement annuel de la cavité peut être estimé en calculant l'intégrale de la courbe des températures souterraine sur un cycle hivernal. L'indice de gel que l'on obtient de cette manière permet une bonne comparaison du refroidissement agissant sur une structure d'une année à l'autre. L'enregistrement des températures de la glacière de Monlési durant l'hiver 2001/2002 présente une valeur de l'ordre de Fi=300°C/jour[1]. Une comparaison entre cette valeur et celles obtenues aux stations météorologiques situées à proximité (par ex. La Brévine ou La Chaux-de-Fonds) permet ainsi de positionner ce cycle dans le contexte climatique des dernières décennies.

Conclusion et perspectives

Les mesures obtenues durant cette première phase de l'étude permettent de cibler les objectifs d'une acquisition plus détaillée. Un pas d'échantillonnage très court est nécessaire pour évaluer la vitesse de refroidissement en période hivernale lors d'événements précis. Une comparaison de l'évolution de l'indice de gel au cours des dernières décennies devrait en outre permettre une meilleure évaluation des paramètres conditionnant la formation/préservation de la glace et donc de donner une indication de la vulnérabilité de la glacière dans le contexte du réchauffement climatique actuel.

Il est probable que les fluctuations climatiques passées sont conservées dans le volume de glace de Monlési. Son analyse permettra de mieux comprendre l'évolution de la glacière jusqu'à ce jour. De cette manière, nous espérons pouvoir identifier le comportement futur de ce remplissage exceptionnel.

Remerciements

Cette étude s'inscrit dans le cadre d'un projet de recherche visant à comprendre la formation et l'évolution des glacières du Jura, soutenu par le Fonds national suisse de la recherche scientifique (projet SNF n°21-63764.00). Nous tenons à remercier R. Delaloye (UNIFR) pour la mise à notre disposition d'une première série de minilogger, utilisés au début de cette acquisition. Merci également aux nombreuses personnes qui nous accompagnent régulièrement sur le terrain dans le cadre de cette étude ou qui nous apportent occasionnellement leur soutien. ■

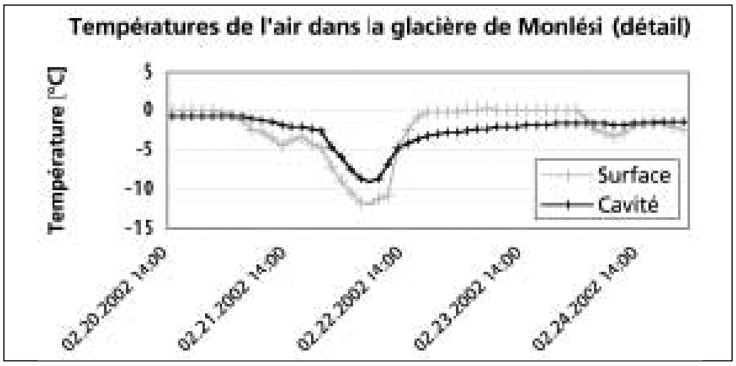

Fig. 2: Le refroidissement dans la glacière n'est que de courte durée. La mise à l'équilibre des températures survient quelques temps seulement après une arrivée d'air froid dans la cavité.

Abb. 2: Die Abkühlung im Innern der Glacière ist nur von kurzer Dauer. Die Einstellung des Gleichgewichtes erfolgt kurz nach dem Eindringen der kalten Luft in die Höhle.

Wetterstationen in der Umgebung (z.B. La Brévine oder La Chaux-de-Fonds) wird es erlauben, den Zyklus im Zusammenhang mit den Klimata der letzten Jahrzehnte zu verstehen.

Schlussfolgerungen und Ausblick

Die Messwerte dieser ersten Phase erlauben es, das Vorgehen für eine detailliertere Untersuchung genauer zu definieren. Insbesondere scheint ein sehr kurzes Messintervall nötig, um die Geschwindigkeit der Abkühlung im Winter bei Temperaturstürzen bestimmen zu können. Im weiteren sollte die Entwicklung der Frostindizes während der letzten Jahrzehnte erlauben, die Parameter, welche für die Bildung und Erhaltung des Eises wichtig sind, besser zu erfassen. Somit kann auch die Empfindlichkeit des Eises im Umfeld der aktuellen Klimaerwärmung abgeschätzt werden.

Im Eisvolumen der Glacière de Monlési sind sicher vergangene Temperaturschwankungen registriert. Eine Eis-Analyse wird uns die Entwicklung des Eisvolumens bis zum heutigen Tag besser verstehen lassen. Damit hoffen wir auch, das zukünftige Verhalten dieser aussergewöhnlichen Eisfüllung verstehen zu können.

Dank

Diese Studie steht im Rahmen eines Forschungsprojektes zum Verständnis der Bildung und Entwicklung von Eis in Höhlen des Juras. Dieses wird vom Schweizerischen Nationalfonds unterstützt (SNF-Projekt Nr. 21-63764.00). Wir danken R. Delaloye (UNIFR) für die Serie von Miniloggern, die für den Beginn dieser Messserie zur Verfügung gestellt wurde. Ebenso danken wir den zahlreichen Personen, die uns für dieses Projekt regelmässig im Gelände begleiteten oder uns gelegentlich unterstützten. ■

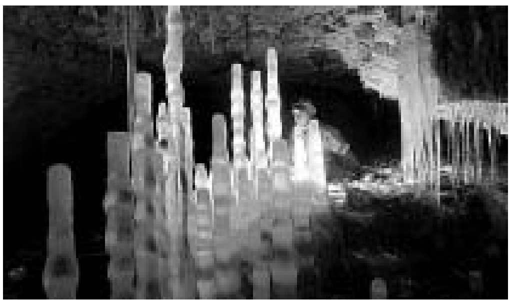

Photo Rémy Wenger

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Nouveau levé topographique de la glacière de Monlési

*par Marc Luetscher et Rémy Wenger (ISSKA)**

La glacière de Monlési, cavité connue de longue date pour son important remplissage de glace, fit l'objet d'un premier croquis par Browne (1865) en complément de ses excellentes observations. Il fallut toutefois attendre près d'un siècle pour voir une première topographie précise de la glacière, utilisable comme support pour une étude plus approfondie. Ce plan, levé par deux étudiants de La Chaux-de-Fonds (Monard & Stettler 1959), permit ainsi une description détaillée de cette cavité et de son remplissage.
Suite à une proposition de classification de la glacière comme géotope national, quelques membres du Spéléo Club des Montagnes Neuchâteloises (SCMN) décidèrent, au printemps 2000, de procéder à un nouveau levé topographique de la cavité. Fort utile pour les études menées dans le cadre d'un projet de recherche en climatologie souterraine, la topographie fut achevée en 2001 par deux collaborateurs de l'Institut suisse de spéléologie et de karstologie (ISSKA).

SITUATION GÉOGRAPHIQUE

Située à la limite du Val-de-Travers et de la vallée de la Brévine, la glacière de Monlési se trouve sur le territoire communal de Boveresse / NE (534'950 /198'925 – 1135 m), à proximité de la ferme des Sagnettes. Cette glacière, protégée par un bosquet d'arbre, s'ouvre au cœur d'un pâturage boisé, typique du paysage jurassien. La topographie locale est caractérisée par la présence d'un petit bassin fermé, qui pourrait favoriser la formation d'un lac d'air froid.

Le climat régional s'illustre par des conditions typiques des hauts plateaux jurassiens. La température moyenne annuelle est de l'ordre de 5 °C alors qu'une pluviométrie annuelle de quelque 1500 mm peut être estimée à partir des relevés de l'Institut suisse de météorologie.

CONTEXTE GÉOLOGIQUE ET HYDROGÉOLOGIQUE

La glacière de Monlési appartient à un alignement de dolines situées dans l'axe du synclinal «Les Parcs – La Glacière – Derrière Le Châble». Elle s'ouvre à la base des calcaires lités du kimméridgien présentant en cet endroit un pendage subhorizontal. La fracturation très dense observée au sein de la glacière permet une infiltration aisée des eaux météoriques. Fortement influencées par la température régnant au sein de la glacière, ces arrivées d'eau favorisent une importante gélifraction contribuant à l'agrandissement de la cavité (Pancza 1992).

Situées dans le bassin d'alimentation de l'Areuse, les eaux souterraines régionales devraient être drainées par le synclinal en direction de la source localisée à quelque 4 km au sud-ouest.

DESCRIPTION

Trois puits d'accès, d'une vingtaine de mètres de profondeur, permettent d'atteindre la salle principale de la glacière de Monlési. D'environ 45 x 25 m, cette salle est partiellement remplie d'une importante masse de glace, estimée à quelque 10'000 m^3 par Stettler & Monard (1960). En bordure de la salle, une importante rimaye permet de longer ce volume de glace sur une dizaine de mètres d'épaisseur jusqu'au point bas de la cavité à -33 m.

Bien que la géométrie des puits favorise l'accumulation de neige en période hivernale, ce remplissage particulier semble essentiellement constitué de glace de regel, issue des nombreuses infiltrations d'eau présentes au sein de la cavité. Ces dernières sont par ailleurs étroitement liées aux quelques cheminées remontant à proximité de la surface.

Les trois puits d'entrée, de diamètres fort différents, favorisent d'importantes circulations d'air par convection au sein de la cavité. Browne (1865) fut ainsi le premier auteur à mettre en évidence une alternance régulière de la direction du courant d'air. Bien que ne pouvant en estimer l'influence, cet auteur attribua une forte importance à ce phénomène dans la genèse du remplissage.

LEVÉ TOPOGRAPHIQUE

Méthode

Le levé topographique de la cavité a pu être réalisé au moyen des méthodes spéléologiques traditionnelles (Grossenbacher 1991). L'utilisation de boussoles et clinomètres optiques de type «Suunto» fut complétée d'un distance-mètre laser Leica.

**Institut Suisse de Spéléologie et de Karstologie; Case postale 818; CH-2300 La Chaux-de-Fonds*

Afin de permettre une meilleure représentation de la surface du terrain au droit de la cavité, une polygonale à été implantée dans la forêt aux alentours de l'entrée. Comportant quelque 37 stations, celle-ci permet, entre autres de positionner un certain nombre de bornes disposées autour de la cavité (cf. Hapka 2002). De même, chacun des puits fit l'objet d'un levé particulier, permettant une meilleure représentation de sa géométrie.

Le choix du cheminement souterrain s'inscrit dans une volonté de cartographier le pourtour de la salle, permettant une meilleure représentation de la répartition spatiale de son remplissage. Une polygonale de 55 stations a ainsi pu être implantée, longeant les parois de la cavité et les puits d'accès. Pour assurer la précision finale, différentes boucles de ces polygonales ont par ailleurs été fermées (fig. 1).

Incertitude de lecture et précisions des mesures

L'utilisation de boussoles et clinomètres « Suunto » implique une incertitude de lecture de ± 1 grade sur chacune des visées. Le distance-mètre laser permet quant à lui d'obtenir une précision de mesure équivalant à ± 1 cm lors d'un levé en cavité. Chaque visée a été mesurée simultanément dans les deux sens. Cette marche à suivre, permettant de supprimer de nombreuses erreurs de lecture, s'est révélée particulièrement efficace en regard de la précision finale obtenue (tableau 1). Il est néanmoins possible qu'une certaine imprécision persiste sur la distance mesurée. En effet, celle-ci ne fut généralement relevée que dans une direction et le risque d'un mauvais positionnement du point de visée existe, d'autant plus fortement que la longueur entre les deux stations est importante.

La précision finale obtenue sur chacune des stations correspond à une moyenne de quelque 5 cm dans les trois directions. L'erreur mesurée, inférieure à 0.5 % de la longueur du cheminement, est nettement plus faible que l'erreur initialement estimée. L'obtention de cette précision met ainsi en évidence le soin apporté au levé de terrain.

Dessin

Afin de disposer d'une topographie suffisamment précise pour bien situer les observations futures, une échelle au 1 : 200e a été définie. Le levé de deux plans distincts, l'un représentant la surface et l'autre la cavité elle-même, s'est révélé souhaitable pour une meilleure compréhension de la glacière (fig. 3).

De plus, trois coupes verticales de la cavité ont pu être levées, recoupant toutes au minimum un puits d'accès (fig. 2). Ces coupes devraient par ailleurs permettre une meilleure représentation ultérieure du volume en trois dimensions.

ESTIMATIONS ET FLUCTUATIONS DU VOLUME DE GLACE

Le levé topographique actuel permet de mieux évaluer le volume de glace contenu dans la glacière de Monlési. En assimilant ce remplissage à un corps rond (segment sphérique à deux bases), il est possible d'estimer un volume de glace de quelque 6'000 m^3. Inférieure de 40 % au volume estimé par Monard et Stettler (1960), cette valeur reste toutefois fortement tributaire de la géométrie choisie pour l'approximation. Une estimation plus rigoureuse nécessite dès lors une meilleure connaissance de la géométrie de l'encaissant. En effet, cette approximation ne repose pour l'heure que sur une estimation visuelle de l'épaisseur du remplissage. Il en ressort une marge d'erreur considérable sur la géométrie de sa base, qu'il ne sera possible de préciser qu'au moyen de sondages.

Malgré cette incertitude, de nombreux témoignages oraux mentionnent une forte régression du volume de glace au cours de cette dernière décennie. Si des mesures régulières de ce volume font encore défaut, les origines diverses de ces observations semblent accréditer la thèse d'importantes fluctuations de la géométrie du glacier souterrain.

Une comparaison des levés topographiques antérieurs (Stettler 1971) et du plan actuel de la cavité indique un récent fluage du glacier vers le sud. Ce changement de géométrie a dès lors engendré l'obstruction complète d'une petite salle située en paroi sud de la salle principale. Cette dernière était encore accessible à la fin des années cinquante, malgré un volume de glace estimé plus important.

Si l'importance de ces fluctuations est encore relativement méconnue, il est toutefois possible d'émettre quelques hypothèses quant à leur origine. En effet, la glacière de Monlési étant essentiellement constituée de glace de regel, il suffit d'un écoulement différent des eaux d'infiltrations pour que la géométrie du remplissage varie. Ce dernier sera en outre directement dépendant des conditions climatiques extérieures et des influences anthropiques sur le milieu souterrain. L'étude en cours devrait dès lors permettre de mieux comprendre l'évolution future d'un tel remplissage.

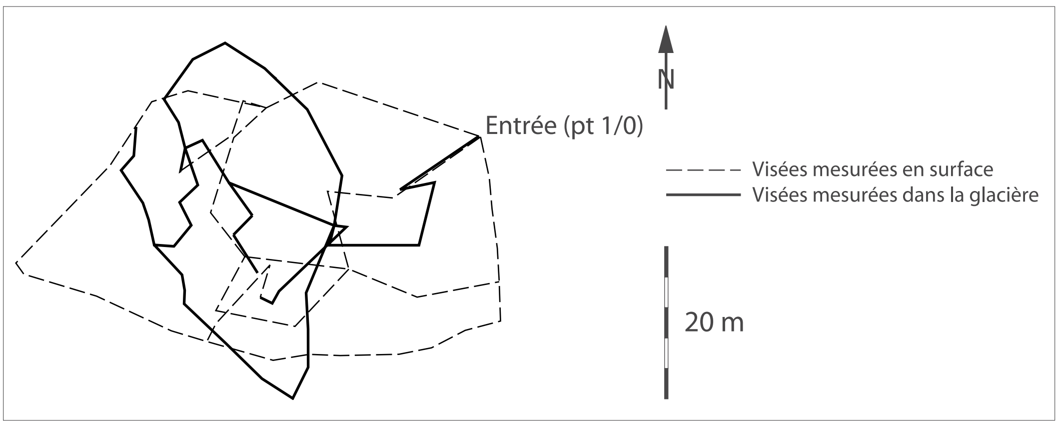

Fig. 1 : Boucles topographiques implantées sur le site de Monlési. La fermeture de nombreuses boucles permet d'obtenir une bonne précision finale sur chacune des stations.

	Segement	Point de départ		Point d'arrivée		nombre de station	Erreur mesurée / erreur estimé			Erreur mesurée [m]			Erreur mesurée par station [m]			Erreur estimée [m]		
	Série	série	point	série	point		Est	Nord	Altitude	Est	Nord	Altitude	Est	Nord	Altitude	Est	Nord	Altitude
SURFACE	1	1	0	1	1	1	0.14	0.14	0.2	0.01	0.01	-0.02	0.01	0.01	0.02	0.09	0.1	0.09
	1	1	1	1	4	3	0.24	0.18	0.3	0.03	-0.03	0.04	0.02	0.02	0.02	0.13	0.17	0.13
	1	1	4	2	2	16	1.61	0.08	0.12	-0.72	0.04	0.06	0.18	0.01	0.02	0.45	0.48	0.52
	2	2	0	2	2	2	0.24	0.34	1.48	-0.04	0.12	-0.55	0.03	0.09	0.39	0.18	0.35	0.37
	20	1	0	20	3	3	0.39	0.37	0.71	-0.1	-0.1	0.22	0.06	0.06	0.13	0.24	0.26	0.31
	20	20	3	1	4	2	0.83	0.28	0.49	-0.13	0.06	-0.11	0.09	0.04	0.08	0.15	0.23	0.23
	24	2	0	1	1	2	0.07	0.45	1.34	0.02	-0.08	0.39	0.01	0.06	0.27	0.27	0.18	0.29
	21	20	3	21	3	3	0.63	1.66	0.56	-0.14	-0.4	-0.15	0.08	0.23	0.09	0.22	0.24	0.27
	22	21	3	2	2	4	0.84	0.49	1.07	0.18	-0.09	0.26	0.09	0.05	0.13	0.22	0.18	0.24
	23	21	3	20	3	1	0.77	0.91	1.43	-0.09	-0.2	-0.31	0.09	0.2	0.31	0.11	0.22	0.22
CAVITÉ	3	3	0	2	0	5	0.29	0.48	0.22	-0.08	-0.15	0.07	0.03	0.07	0.03	0.27	0.31	0.33
	9	9	1	9	2	1	0.01	0.1	0.01	0	-0.01	0	0	0.01	0	0.1	0.1	0.11
	9	9	2	3	0	1	0.01	0.21	0.01	0	-0.05	0	0	0.05	0	0.13	0.22	0.24
	8	3	0	7	5	10	0.96	0.33	1.23	0.27	0.1	-0.37	0.09	0.03	0.12	0.28	0.31	0.3
	7	7	5	6	8	6	0.77	0.2	0.75	0.18	0.04	-0.14	0.07	0.01	0.06	0.23	0.19	0.18
	9	6	8	9	1	1	0.01	0.05	0	0	0	0	0	0	0	0.09	0.05	0.04
	4	2	2	4	4	4	0.29	0.34	0.13	-0.08	-0.07	0.03	0.04	0.04	0.01	0.27	0.22	0.2
	4	4	4	6	8	1	0.05	0.15	0.03	0	-0.01	0	0	0.01	0	0.05	0.1	0.05
	6	3	0	6	3	3	0.46	0.09	0.78	-0.1	-0.02	0.17	0.06	0.01	0.1	0.21	0.19	0.22
	6	6	3	6	8	5	0.43	0.09	0.64	-0.08	-0.02	0.12	0.04	0.01	0.05	0.19	0.21	0.18

◄ *Tabl. 1 : Tableau des erreurs de bouclages. Les valeurs indiquent une précision de mesure meilleure que celle initialement estimée (1 grade sur les instruments de visée et 0.1m sur la longueur).*

CONCLUSION ET TRAVAUX FUTURS

Cette nouvelle topographie permet de disposer d'un levé actuel et de bonne précision de la glacière de Monlési. La mise en évidence d'importantes modifications de la géométrie du remplissage au cours des cinquante dernières années encourage une étude détaillée de ce remplissage. Dans ce but, divers repères ont été implantés afin d'obtenir des mesures objectives de ces variations. Par ailleurs, dans l'objectif d'évaluer plus justement le volume de glace présent, différents sondages sont prévus pour mesurer l'épaisseur de la glace. Cette étape devrait dès lors contribuer significativement à une meilleure compréhension de l'évolution de ce remplissage particulier.

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Tous nos remerciements s'adressent également aux deux stagiaires de l'ISSKA, C. Guillez et F. Hof pour leur important soutien sur le terrain. Enfin, nous ne saurions oublier le Fonds National (projet SNF n°21-63764.00) sans qui l'étude, dans laquelle s'est inscrit ce levé, ne pourrait se faire.

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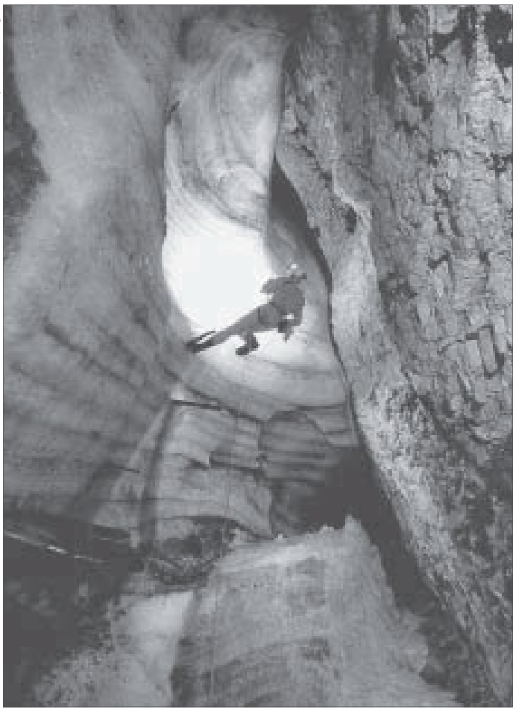
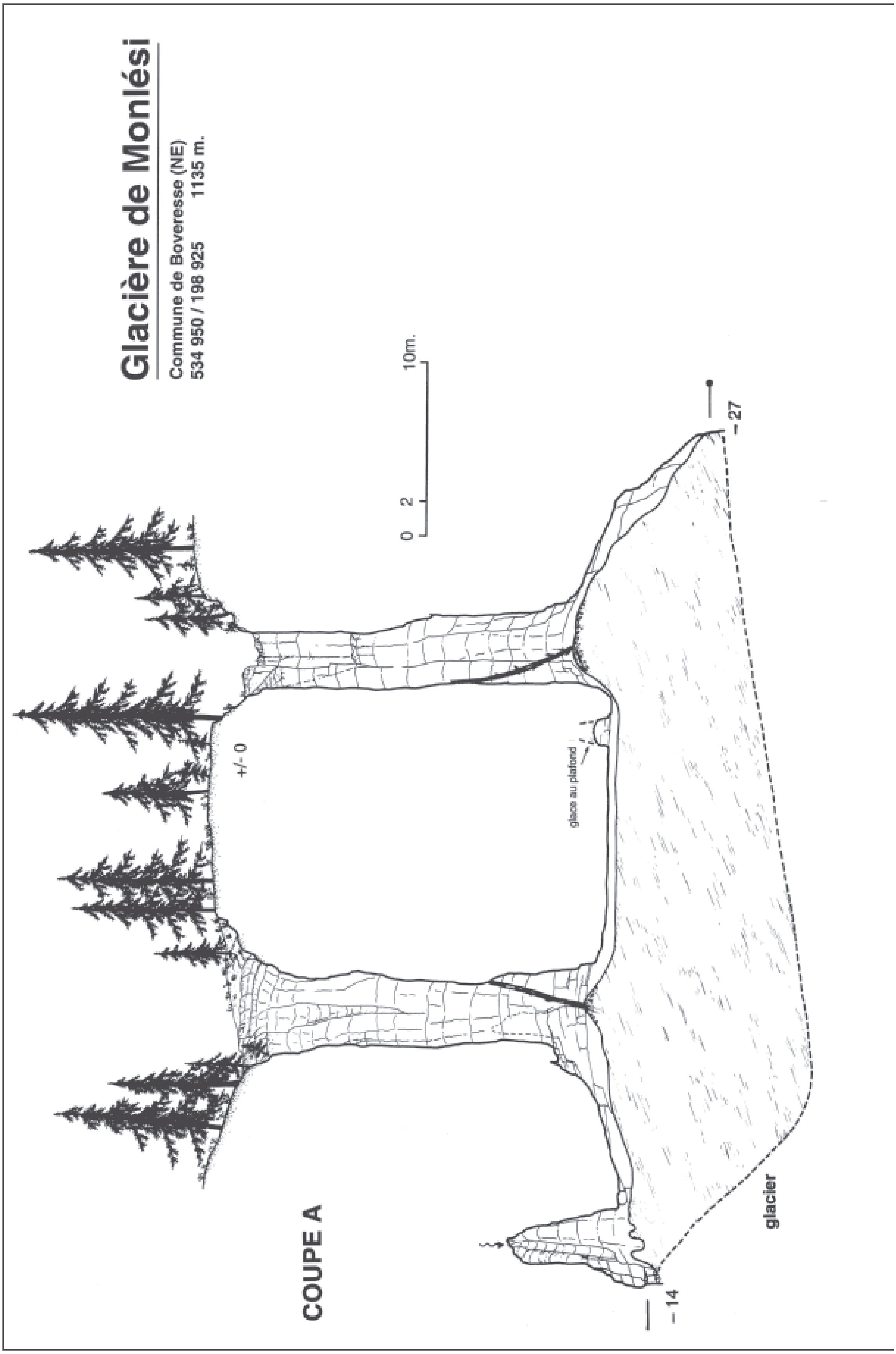
►► *Fig. 2 : Coupes verticales de la glacière de Monlési (échelle original 1 : 200e) La coupe A est parallèle au grand axe de la salle alors que les coupes B et C lui sont perpendiculaires).*

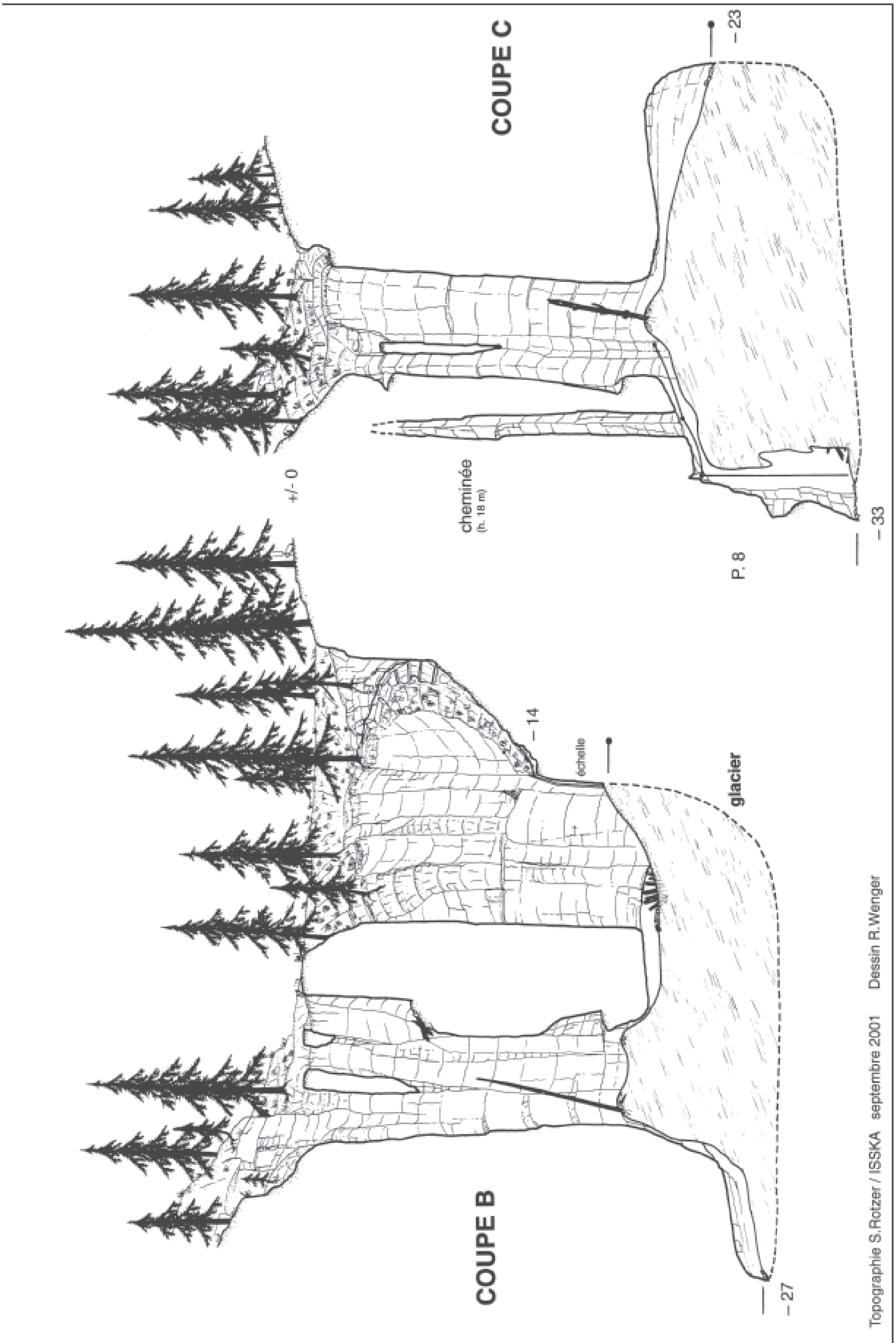
►►►► *Fig. 3 : Plan de la glacière de Monlési et de la surface avoisinant les puits d'entrées (échelle original 1 : 200e).*

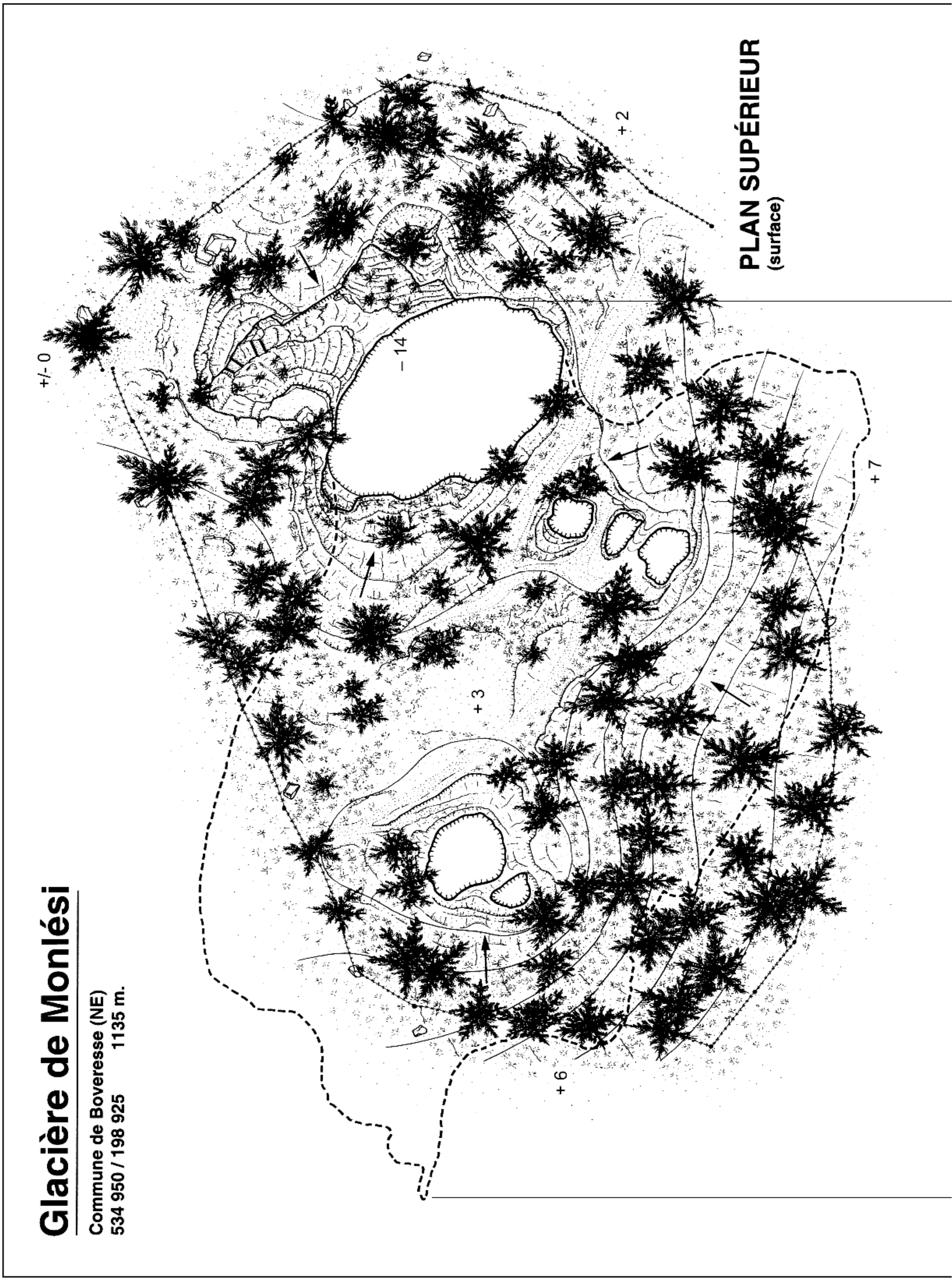
photo R. Wenger



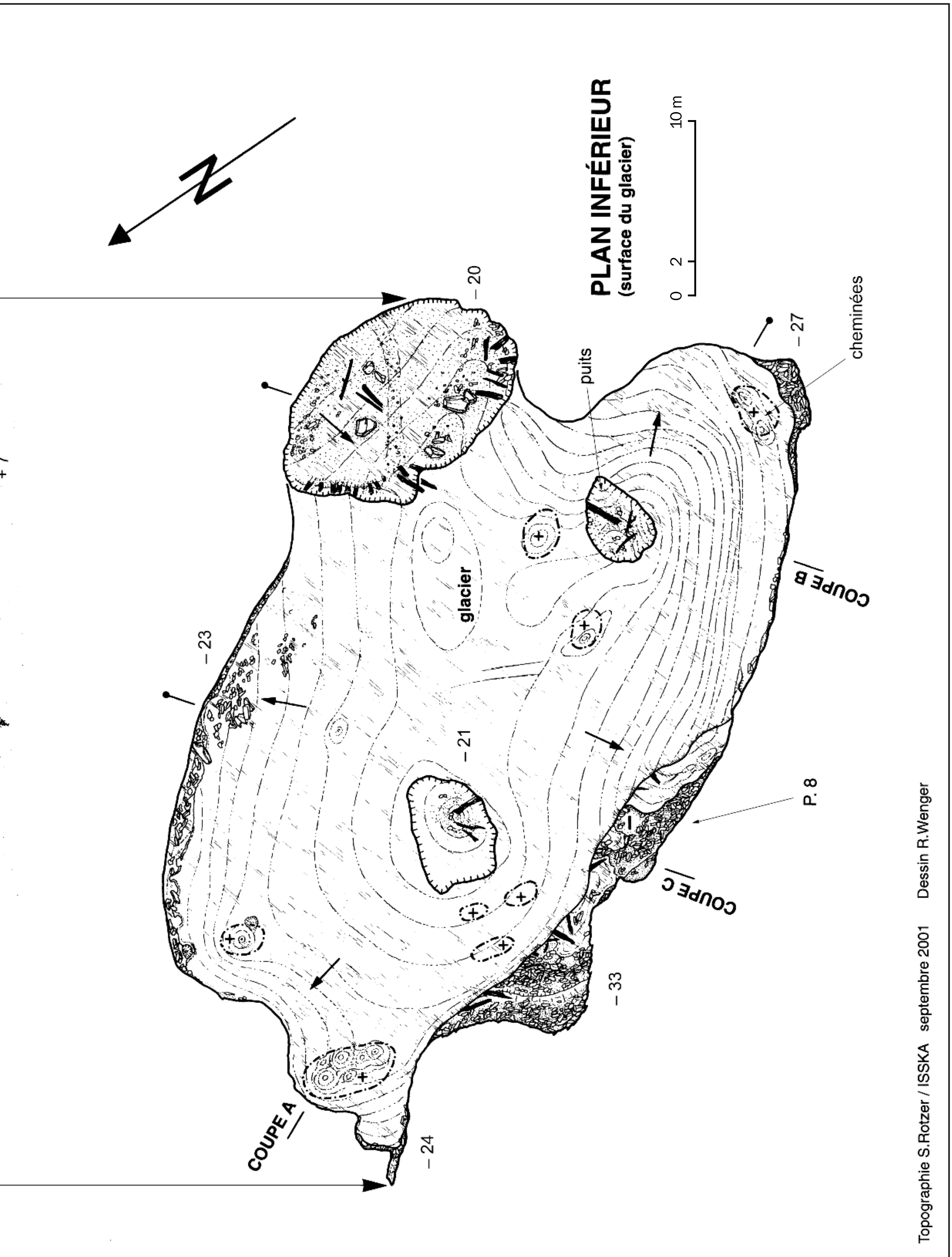
◄ *Photo 1 : Une importante rimaye permet de longer le remplissage de glace sur son flanc ouest sur quelque 8 m d'épaisseur.*


Glacière de Monlési
Commune de Boveresse (NE)
534 950 / 198 925 1135 m.
0 2 10m.
COUPE A
+/- 0
glace au plafond
glacier
– 14
– 27


COUPE C
– 23
+/- 0
cheminée
(h. 18 m)
P. 8
– 33
– 14
échelle
glacier
COUPE B
– 27
Topographie S.Rotzer / ISSKA septembre 2001 Dessin R.Wenger



Glacière de Monlési
Commune de Boveresse (NE)
534 950 / 198 925 1135 m.
PLAN SUPÉRIEUR
(surface)
+/- 0
+ 2
–14
+ 7
+ 3
+ 6


N
PLAN INFÉRIEUR
(surface du glacier)
0 2 10 m
– 20
– 27
cheminées
puits
glacier
– 23
– 21
COUPE B
P. 8
COUPE C
– 33
COUPE A
– 24
Topographie S.Rotzer / ISSKA septembre 2001 Dessin R.Wenger

PART III:

PROCESSES IN ICE CAVES

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The presence of subsurface ice is documented in numerous alpine karst regions even though the mean annual air temperature sometimes greatly exceeds 0 °C. Despite the historical interest raised by these «ice caves», little data is available to predict their evolution in a changing climate context. This doctoral research project approaches the issue from a thermodynamics perspective. Data from a case study of Monlési ice cave in Switzerland emphasizes the major role of subsurface air circulations in the formation and the conservation of cave ice. Results lead to the conclusion that the cave ice mass balance reflects mainly the external winter atmospheric conditions. The dating of thousand-year-old ice deposits predicts interesting paleoclimatological contributions.

La présence de glace souterraine est documentée dans la plupart des karsts alpins, et ce bien que la température moyenne de l'air extérieur dépasse parfois largement le 0 °C. Malgré l'intérêt historique suscité par ces «glacières», peu d'éléments permettent de prédire leur évolution dans un contexte climatique changeant. Ce travail de doctorat aborde la question sous l'angle de la thermodynamique. Une étude de cas menée sur la glacière de Monlési (Suisse) souligne le rôle majeur des circulations d'air hypogées pour la formation et la conservation de la glace souterraine. On y conclut que le bilan de masse d'une glacière reflète principalement les conditions atmosphériques hivernales extérieures. La datation de dépôts de glace millénaires présage d'intéressantes contributions paléoclimatologiques.

Unterirdisches Eis wird in den meisten alpinen Karstregionen dokumentiert und dies, obwohl die durchschnittliche Aussenlufttemperatur die Nullgradgrenze manchmal weit überschreitet. Trotz des historischen Interesses für diese «Eishöhlen», sind nur wenige Daten vorhanden um deren Entwicklung in einem ändernden Klimakontext vorauszusagen. Diese Doktorarbeit behandelt die Frage unter dem Aspekt der Thermodynamik. Eine Fallstudie aus der Eishöhle Monlési (Schweiz) unterstreicht die Hauptrolle der unterirdischen Luftzirkulationen für die Bildung und die Erhaltung des Höhleneises. Daraus folgt, dass die Massenbilanz einer Eishöhle hauptsächlich die winterlichen Aussenbedingungen widerspiegelt. Die Datierung tausend jährigen Eisablagerungen kündigt interessante paläoklimatische Beiträge an.